Structural development of the Håsteinen Devonian Massif, its Caledonian substrate and the subjacent Nordfjord–Sogn Detachment Zone — a contribution to the understanding of Caledonian contraction and Devonian extension in West Norway

Volume I

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Dissertation for the degree doctor scientiarum (dr.scient.)
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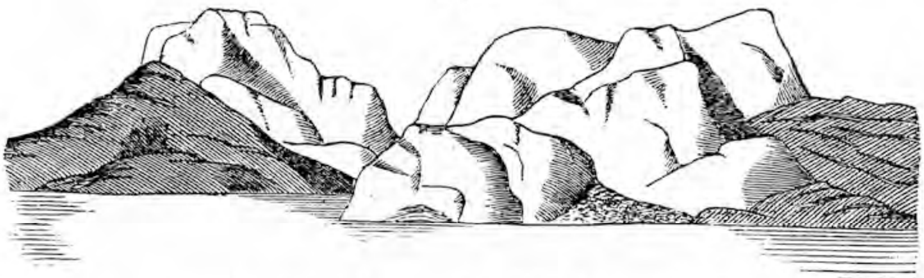
April 2008

Conglomeratet ligger i Bunden af Høgdalsfjorden over Lerglimmerskiferen og danner det høje klumpede Fjeld Haastenen. Landskabet i Bunden af Høgdalsfjorden viser et af de tydeligste og mest storartede naturlige Profiler, man nogetsteds kan se, selv paa denne paa interessante Fjeldformer saa rige Kyst. Over det svagt skraanende, mørkt farvede og skovbevoxede Lerglimmerskiferfjeld, hæver sig brat og klumpformig Haastenen med nøgne ensformig graa Vægge. Fig. 5 forsøger i grove Træk at skizzere denne mærkelige Fjeldform.

Fig. 5.



Allernederst optræder hist og her en ubetydelig Lagfølge af rød Sandstenskifer; ellers bestaar hele Massen af Conglomerat. Runde og skarpkantede Brudstykker af alle Størrelser, hovedsagelig bestaaende af forskjellige Kvartsiter og Kvartsskifere, ligge i stor Mængde ved Siden af hverandre, saa at Bindemidlet aldeles mangler. Hvor Brudstykkerne ere færre, er Bindemidlet sandstenagtig med mange smaa Kvartskorn.

Irgens & Hiortdahl, 1864

ABSTRACT

In western Norway, the contractional history of the Caledonian Orogeny ended with the Scandian continent-continent collision between Baltica and Laurentia at around **420-400 Ma**, resulting in formation of eclogites in the subducted Baltic margin (**Western Gneiss Region, WGR**) and the top-E emplacement of outboard terreranes (for example the **Høydalsfjorden Complex, HC**) and nappes of Baltic affinity. The collapse of the orogenic belt started at the end of the Early Devonian period, with reverse top-W movement of the entire Scandian nappe pile on the shallowly W-dipping basal decollement zone. This movement, termed **Mode-I**, was dated to **402–394 Ma** (Fossen & Dunlap 1998), and interpreted to mark the start of divergent movements between the Laurentian and Baltic plates. During the Mode-I movements, the eclogitic WGR experienced exhumation together with the above-lying Mode-I zone. At the end of the Mode-I movements, the Mode-I zone had lost its westward dip, and the resulting locking of the zone led to formation of the **Mode-II Nordfjord-Sogn Detachment zone (NSDZ)**, cutting through the Scandian nappe pile. The NSDZ zone was a shallowly W-dipping mega-shear zone that divided the orogenic belt into an eastern "Lower Plate" and a western "Upper Plate". The Upper Plate moved down- and W-wards on the NSDZ, causing formation of basins that were filled with continental Devonian clastic sediments.

In the study area, located **15 km ESE** of the town of Florø in the municipality Flora, Sogn og Fjordane, three tectonostratigraphic units are present. These are from base to top, the **Eikefjord Group (EG)**, the **Høydalsfjorden Complex (HC)** and the **Håsteinen Devonian Massif (HDM)**. Each of the units displays *different* tectonometamorphic histories. The EG is penetratively mylonitised and interpreted to be part of the NSDZ. The HC is part of the above-lying Upper Plate, and the sediments of the HDM were deposited on top of the HC.

Precambrian/early Caledonian geological history

The earliest geological history in the study area can be found in the Eikefjord Group, which contains Precambrian orthogneisses dated to **1511 +/- 64 Ma**, interpreted as a minimum intrusive age (Abdel-Monem & Bryhni 1978). The rocks consist of meta-anorthosite, grey monzonitic biotite-epidote gneiss, amphibolites, meta-gabbro and minor micaschist, and have been assigned to the "anorthosite-Jotun Kindred" which is correlated with the Jotun and Dalsfjord Nappes. Remnants of an amphibolite facies mineralogy that define the highest metamorphic grade in the rocks is thought to be Caledonian or post-Caledonian.

Caledonian (Scandian) contractional geological history

In the **Eikefjord Group (EG)**, the only possible relict Scandian feature is the amphibolite facies mineralogy that may have formed during the subduction of the WGR. The mineralogy has been strongly retrograded during the post-Scandian NSDZ. The **Høydalsfjorden Complex (HC)** displays the Scandian (top-E) development of the rocks in the study area. The complex essentially consists of gabbro-intruded metasediments with well preserved bedding (**S_0**). The rocks are either the cover sequence to an ophiolite and/or a melange and may be related to the **443 +/- 4 Ma** Solund–Stavfjorden Ophiolite Complex (Dunning & Pedersen 1988). Obduction onto the Baltoscandian platform occurred in the time interval **425–400 Ma**, yielding a bedding-parallel **S_1**-foliation with a prograde lower greenschist facies (upper chlorite-grade) **M_1**-metamorphism. The **S_1**-fabric was folded into W-E to WNW-ESE trending **F_2**-folds. **S_2**-fabrics are represented by kink bands and locally by a crenulation cleavage that displays only vague recrystallisation, with metamorphism still corresponding to lower greenschist facies (upper chlorite grade). The **F_2**-folds are cut by the sub-Devonian unconformity, and the **D_1**- and **D_2**-deformations are thus pre-dating deposition of Devonian sediments.

Post-Caledonian, Devonian, extensional geological history

General: The subdivision of the Devonian extensional phase into Mode-I movements and the subsequent NSDZ/Mode-II structures is reviewed above. The extensional tectonics produced two tectonic styles in the field area: first the onset of top-W extensional movements, then the onset of folding.

Mismatch: The present thesis reveals that a discrepancy appears to exist between the $^{40}Ar/^{39}Ar$ muscovite ages of the NSDZ in the Eikefjord–Gloppen area and the status of the NSDZ as a Mode-II zone. The $^{40}Ar/^{39}Ar$ muscovite ages in the Eikefjord–Gloppen area fall in the range **404–398 Ma** (Chauvet & Dallmeyer 1992; Andersen 1998), which is contemporaneous with the movements at **402–394 Ma** on the Mode-I zone (Fossen &

Dunlap 1998). In addition, the NSDZ shows extensive movements at amphibolite facies conditions, indicating that the movements on the NSDZ started even earlier than the time interval **404–398 Ma**. Recently, Johnston et al. (2007b) performed P/T analyses and radiometric dating in the Eikefjord area, and interpreted the NSDZ to have moved in the time interval **410–400 Ma**. Viewed together, these ages may seem to indicate that the NSDZ initiated *before* the Mode-I zone, i.e. opposite to the conventional order of appearance. However, as the order of the "Mode-I first" and "Mode-II next" in this area is based on independent field work (Milnes et al. 1997), the conventional order of appearance has been maintained. The cause(s) of the discrepancy is not clear.

Effects of Devonian extensional movements on the Høydalsfjorden Complex (top-W movements predating deposition of the HDM). The post-Scandian extensional movements led to formation of top-W $\mathbf{S_3}$-mylonites in the Høydalsfjorden Complex (HC). A very limited number of folds with NE-SW trending fold axes has been recorded, i.e. oriented with a high angle to the abundant WNW-ESE trending HC $\mathbf{F_2}$-folds, the latter cut by the unconformity below the Håsteinen deposits. The NE-SW trending folds may be $\mathbf{F_3}$-folds related to the $\mathbf{D_3}$-top-W mylonites that are cut by the unconformity below the Devonian Håsteinen sediments. This suggests that the $\mathbf{S_3}$-mylonites formed at an early stage of the Devonian extension, *prior* to the Devonian basin formation. It appears that the HC $\mathbf{S_3}$-mylonites are unrelated to the Eikefjord Group mylonites of the **Mode-II**/NSDZ. This is consistent with the lower metamorphic grade in the HC $\mathbf{S_3}$-mylonites, where brittle quartz suggests temperatures *below* greenschist facies. Instead, the HC $\mathbf{S_3}$-mylonites may be related to the foregoing **Mode-I** extension.

Effecs of Devonian extensional movement on the Eikefjord Group. The top-W $\mathbf{D_2}$-movements of the NSDZ produced the strong and penetrative mylonitic $\mathbf{S_2}$-fabric that generally obliterates all older structures and is retrogressive from amphibolite facies (mentioned above) down to middle or lower greenschist facies. The amphibolite facies might originally have formed as a result of partly "down-transport" of the EG rocks during subduction of the WGR. The metamorphic retrogression of the EG reflects the uplift of the rocks during the movements on the NSDZ. The related $\mathbf{L_2}$-stretching lineations have a WNW-ESE trend and shallow plunges.

Formation of the Håsteinen Devonian basin. The sediments were accumulated unconformably on top of the HC Upper Plate $\mathbf{D_1}$- and $\mathbf{D_2}$-structures (top-E) as well the $\mathbf{D_3}$-mylonites (top-W). The sediments consist of continental conglomerates deposited by mass flow processes in a proximal alluvial fan environment, and minor sandstones along the westernmost margin, deposited by fluvial processes in a more central or distal part of the fan environment.

Along an E-W section through the middle of the massif, the beds are constantly standing with an eastward **53°** dip against the subjacent, subhorisontal unconformable surface of the Høydalsfjorden rocks. This means that the Devonian beds – after having been deposited *sub-horizontally* – were rotated during the extension to achieve a *steep easterly dip*. The basin floor, defined by the Upper Plate (HC), must have rotated accordingly. The *present-day* **53°** angle between the steeply dipping bedding and the overall subhorisontal contact towards the substrate represents an extraordinary geometry. This angle must have been high also at the *time of sedimentation*, but then with a sub-*horizontal* Devonian bedding deposited against a *W-dipping* palaeo-slope. This geometry cannot be explained by hitherto published models, which suggest that the sediments were deposited directly against the fault plane of a constantly active (listric?) growth fault, implying that the bedding/substrate contact being a tectonic fault rather than a primary unconformity as in the HDM. Hence a new model – termed the **ramp basin model** – is presented in the present work. The unique feature of this model is that the deposition is believed to have occurred in a depression that formed on top of the Upper Plate as a result of a westward-dipping frontal ramp in the subjacent detachment zone. In contrast, previous models suggest that the deposition occurred in a half-graben type basin. In the ramp basin model setting, a traditional half-graben basin *would* form at a considerable distance *east* of the "ramp basin", at a position above the listric detachment zone present near the cut-off line. The Hornelen Devonian deposits appear to have formed in a half-graben basin of the latter type.

The ramp basin model has been tested by **numerical forward modelling**. The modelling has been carried out by means of the structural modelling program **2D-Move**. In the model, horizontal beds are deposited in the ramp depression, with an unconformable contact against the W-dipping surface of the HC. As the W-dipping Upper Plate moves W-wards down the ramp, the Devonian beds progressively rotate to obtain eastward dips. When the Upper Plate and attached beds have reached the flatter part of the NSDZ west of the ramp, the Devonian beds have achieved a prominent E-ward dip, dipping against the subhorisontal Upper Plate, just as in the Håsteinen case. Hence, the bedding-normal, cumulative stratigraphic thickness of **5.8 km** for the HDM, as calculated along the E-W trending axial trace of the Osstrupen syncline, is reproduced. As the purpose of the modelling has been to test whether the geometry of Håsteinen can be reproduced, a full testing of various parameters has not been necessary to carry out.

Devonian folding in general: Folding of Devonian age has affected all three units in the study area. In all the units, the fold axes are trending WNW-ESE, indicating roughly NNE-SSW directed contraction. In the literature, folding of the Devonian deposits and subjacent units in Western Norway have been interpreted as a result of a regional contractional event, affecting the entire area between Bergen and Trøndelag. In the present thesis, however, it is suggested that the folding of the Håsteinen basin and its substrate has mainly been a result of intrabasinal processes related to the extension, indicating that regional N-S contraction may not be necessary to explain the folding.

Devonian folding of the Håsteinen Devonian Massif (HDM). In the HDM, the Devonian folding led to formation of the Osstrupen $\mathbf{F_1}$-syncline; an upright, plane fold with an axial trace oriented WNW-ESE (**295°–115°**), and with a average fold axis with a trend/plunge of **115/53**. The limbs are straight, with a mean strike/dip orientation of **070/62 SE** for the northern limb and **161/62 NE** for the southern limb, and an interlimb angle of **103°**. In the Gravaneset sandstone unit in the far west, a set of parasitic $\mathbf{F_1}$-folds with related axial planar $\mathbf{S_1}$-cleavage have been developed, showing that the major part of the deformation occurred during semiductile/ductile (plastic) conditions. In the thesis, three models have been presented to explain the folding in the HDM: (i) folding due to transpression along converging lateral ramps; (ii) folding due to ridge-shaped frontal ramps ("Ridge-ramp" model). (iii) folding due to narrowing basin at depth ("Narrowing basin depth" model). Of these, model (iii) appears to offer the most likely explanation.

Devonian folding in the Høydalsfjorden Complex (HC). In the Høydalsfjorden Complex, the effects of Devonian N-S folding are less obvious. Both the HC $\mathbf{F_2}$-fold structures that trend roughly parallel to the $\mathbf{F_1}$-fold axis of the HDM Osstrupen syncline, and the top-to-the-west $\mathbf{S_3}$-mylonites, are cut by the sub-Devonian unconformity and thus predate the folding of the HDM. A small number of fold axes that are trending NE-SW, i.e. at a high angle to the $\mathbf{F_2}$-folds, are possibly $\mathbf{F_3}$-folds, although their relationship to the $\mathbf{S_3}$-mylonites and the unconformity have not been observed. $\mathbf{S_3}$-mylonite fabrics may possibly be folded by WNW-ESE trending $\mathbf{F_4}$-fold axes that may have formed during the N-S contraction that also folded the Håsteinen deposits. Generally, it is difficult to reveal the possible effects that the N-S contraction of the HDM had on the HC rocks. Possibly, HC-$\mathbf{F_2}$-folds became tightened during the Håsteinen contraction, or the HC may actually contains Devonian folds that have been erroneously recorded as HC-$\mathbf{F_2}$-folds due parallelism with the latter folds. The only place where Devonian folding can be seen to clearly affect HC rocks is at Gravaneset in the far west, where the unconformity below the Gravaneset Sandstone Unit is folded along with the sandstone beds.

Devonian folding in the Eikefjord Group (EG). In the Eikefjord Group, the Devonian contraction folded the $\mathbf{S_2}$/$\mathbf{L_2}$-mylonite fabric into $\mathbf{F_3}$-folds with fold axes having a WNW-ESE oriented trend and shallow plunge, making the $\mathbf{L_2}$-stretching lineations and the $\mathbf{F_3}$-fold axes parallel. An $\mathbf{S_3}$-axial planar fabric was not developed. Furthermore, the trend of the $\mathbf{F_3}$-folds in the Eikefjord Group is parallel to the trend of the Osstrupen $\mathbf{F_1}$-syncline in the HDM. The Devonian N-S contraction produced the $\mathbf{F_3}$-folding of the $\mathbf{S_2}$-mylonites of the Eikefjord Group, since the $\mathbf{F_3}$-folds are folding the extensional $\mathbf{S_2}$-mylonites. The parallel orientation of $\mathbf{L_2}$-stretching lineation and $\mathbf{F_3}$-fold axes in the Eikefjord Group might indicate a genetic relationship, although this is still somewhat uncertain. Since the $\mathbf{F_3}$-folding in the Eikefjord Group appears to be more intense than the $\mathbf{F_1}$-folding giving the Osstrupen syncline, the $\mathbf{F_3}$-folding could be a result of shear-related contraction within the detachment zone.

Post-Devonian geological history

A large number of faults and joints are developed in the HDM. The amount of displacement has been tested for the most prominent faults crossing the massif, and the movements are found to be negligible. A minor fault at the Devonian-substrate contact beneath the Gravaneset sandstone unit has been dated by palaeo-magnetism to a Triassic/Early Jurassic age (Torsvik et al. 1987).

ACKNOWLEDGEMENTS

The present dr.study has been carried out at the Department of Earth Science (Institutt for geovitenskap), University of Bergen, Norway.

Financial support for the field work was received from Elf Aquitaine Norge a.s. and the University of Bergen.

The following persons and institutions have in various ways contributed to the present study:

Professor Haakon Fossen was my supervisor during the study. He provided great professional, personal and logistical support that were of invaluable significance for the project. He has also been a great sparring-partner in discussions on Caledonian and post-Caledonian geology.

Professor Olav Eldholm, head of department, provided facilities necessary for the work to be completed.

Dr. Heidi Espedal, Head of administration, provided moral and logistical support during the last stages of the work.

Professor Alan G. Milnes was my supervisor duing the first year of the dr.study. He gave the project a very good start before he was called to a new position at the Department of Earth Sciences, Uppsala University, Sweeden.

The late professor Brian A. Sturt originally suggested this challenging area for my study.

During their visits to the field area, several issues were discussed with Professor Harald Furnes, professor Chris Stillman, professor Donald Ramsay, professor Hanan Kisch, Dag Bering, Reidulv Bøe, and John Boile.

Professor Harald Walderhaug in Bergen contributed in several discussions where he sheared his extensive knowledge on palaeomagnetism. He also helped with the construction of Fig. 2.9b.

Professor Arild Andresen, Department of Geosciences, University of Oslo, arranged for me to stay the first year of my dr.study at the department in Oslo. Staff and students there contributed to make it a great experience, professionally and socially.

Professor Roy Gabrielsen allowed me to work in the structural lab at the department in Bergen.

Dr. Øystein Larsen, who worked on the Hornelen Devonian deposits during his dr.scient. study, was a valuable partner in discussions on the geology of Western Norway.

Inge Bryhni, senior curator at the Museum of natural history, Section of Mineralogy and Geology, University of Oslo, provided unpublished maps and structural data on Sunnfjord.

Ole Lutro, researcher at the Norwegian Geological Survey, provided preliminary editions of the geological 1:50.000 maps "Eikefjord" and "Naustdal", as well as 1:40 000 aerial photographs of the Håsteinen area.

Helge Stenseth and the late Bjarne Stavenes prepared the more than 200 thin sections.

Jane Ellingsen offered advice on map drawing and assisted with the final completion of the maps.

Part of the work on the figure volume was outsourced to Allkopi A/S. Ingunn Pedersen made valuable improvements on the large formats (map plates and road logs). Frode Jakobsen had a good hand on the printing of the large formats. Sigbjørn Kronenberger greatly improved the figures on the A4 format and prepared the figure volume. He also carried out the final printing of the thesis. Ivar Hallem, marketing director, superintended the whole process and offered excellent service.

Housing during the field work was provided by the Engebø family in Store Høydal, and Lilly Vasset in Eikefjord.

Generally, students and staff at the department in Bergen participated in numerous geological discussions and created a good atmosphere.

Sturle Trovik gave invaluable moral support and sorted out logistical problems at critical stages of the work. En klippe.

My present employer, Odin Petroleum AS, kindly granted a leave of absence which allowed me to work full time at the institute in Bergen, to complete and submit the thesis.

My family has provided tremendous moral support and encouragement during all stages of the work.

— To all these persons and institutions I express my sincere thanks.

Bergen, april 2008

Vegard V. Vetti

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CHAPTER 1 INTRODUCTION

The present study deals with the geological development of the Håsteinen Devonian Massif and its substrate. The study area is located **15 km** east-southeast of the town of Florø in Flora kommune (municipality), in the Sunnfjord region of Sogn og Fjordane fylke (county), western Norway. The present work has been submitted as a dissertation for the degree of Doctor Scientiarum (Ph.D.) at the Department of Earth Science, University of Bergen.

The studied area contains three major tectonostratigraphic/stratigraphic units. These are from top to bottom:

(1) **Håsteinen Devonian Massif**, which contains folded "Old Red Sandstone" conglomerates and minor sandstones of assumed Middle Devonian age. (Age inferred from correlations with fossil-bearing neighbouring Devonian massifs). In the study area, the Håsteinen massif rests with a tectonically modified primary unconformity on the

(2) **Høydalsfjorden Complex**, which contains meta-sediments and intrusive meta-gabbros of assumed Lower Palaeozoic age (possibly from the Ordovician-Silurian boundary; age inferred from lithologic correlations with the Solund–Stavfjorden Ophiolite Complex and associated rocks, previously radiometrically dated to **443 ± 3 Ma**). The Høydalsfjorden Complex is in fault contact with, but tectonostratigraphically on top of the

(3) **Eikefjord Group**, which contains meta-anorthosites and other related meta-igneous rocks (gneisses) of Precambrian age (previously radiometrically dated to **1511 ± 64 Ma**); which are strongly affected by penetrative top-to-the-west mylonitization of late-/post-Caledonian age. The Eikefjord Group apparently constitutes part of the Nordfjord–Sogn Detachment Zone.

1.1 PROJECT DESCRIPTION

Aims of the project

The present work was started as a cand. scient. thesis (= master thesis), but prior to graduation converted to a doctor scient. thesis (= PhD thesis).

The study area of the project has been the *western half* of the Håsteinen Devonian Massif with adjacent rocks. The overall scope has been to identify and describe the different rock units and to deduce the geological history of the area. This was to be accomplished by detailed field mapping of lithologies and structural elements, as well as studies of the tectonometamorphism and aspects of the sedimentology. In analyses of metamorphism, descriptions were to be based on thin-section microscopy; and in analyses of deformation, descriptions were to be based on stereographic plots of structural elements. Numerical modelling was to be applied, to see if certain bedding geometries related to the formation of the Håsteinen and Hornelen basins could be reproduced, thererby also providing a basis for comparison of the two basins.

In more detail, the aims of the study were:

- to map the lithologies and the structural elements on field maps at a scale of 1:5000, and in more detail where desirable.
- in the Eikefjord Group; to analyse deformation and metamorphism, and to describe representative lithologies as a basis for this analysis. Lithologies were not to be differentiated on the map.
- in the Høydalsfjorden Complex; to describe and map lithologies, and to analyse deformation and metamorphism.
- in the Håsteinen Devonian Massif; to map the massif (western half), and to describe the structural geology, the metamorphism, and general aspects of the sedimentology (such as distribution of lithofacies, formations, etc).
- to investigate the contact-relationships between these three major units, particularly "basement-cover" relationships between the Devonian massif and its substrate.
- to present models for the geological development of the area. The numerical modelling program “2Dmove” was to be applied to see if simple models could be made that would reproduce bedding-dips and bedding/substrate angles that formed during the formation of the Håsteinen and Hornelen basins, also allowing comparisons to be made between the two basins. A more comprehensive numerical analysis was beyond the scope of the thesis.

The Devonian Massif of Håsteinen, and its immediate substrate, have not been subjected to separate and detailed geological investigations since the essentially petrographical descriptions of Kolderup (1925). Håsteinen is therefore the last of the major west-Norwegian Devonian massifs to be subjected to modern geological investigations. From the mid 1980s, the formation as well as the deformation of the Devonian basins have become the focus of considerable new interest, as a result of new models which view the basins as "collapse-basins" that formed above crustal-scale detachments due to post-orogenic collapse of the Caledonian mountain chain (e.g. Seranne et al. 1989; Fossen 1992b, 1998, 2000; Krabbendam & Dewey 1998; Osmundsen & Andersen 2001). Very little geological information of relevance to these models has been available from the Håsteinen Devonian Massif and its immediate substrate, and hence the present thesis contributes to fill this gap in the record of Devonian rocks in Norway.

In the light of the general absence of other than petrographical information on the Håsteinen Devonian Massif, it has been considered particularly important to give a presentation of the structure and deformation of the Devonian rocks, with a more limited account of their sedimentology. The investigation of the Lower Palaeozoic rocks of the Høydalsfjorden Complex, which form the immediate substrate to the Devonian massif in the study area, has been carried out in order to separate out – if possible – the pre-Devonian (Scandian?) contractional tectonometamorphic development, from possible later, Devonian overprinting effects. The Eikefjord Group has been investigated to reveal the structural and metamorphic nature of the mylonites, and to find out whether the mylonites are part of the Nordfjord–Sogn Detachment Zone (NSDZ) or not. The NSDZ is important because of its potential role in controlling the formation of the Devonian basins.

The present work therefore deals with a wide range of topics, and it has not been possible to go as deeply into each subject as one could have wished. It is, however, the author's hope, in the future to be able to continue the work on the various fascinating geological challenges of the Håsteinen Devonian Massif and its substrate, and in particular to unravel the details of the easten part of the massif which has not been covered in the present study.

Study area

The study area covers the *western half* of the Håsteinen massif, with adjacent rocks. The area is located west of lake Vassetvatnet (**Fig. 1.1**, **Plate 1**).

The size of the study area is about **60 km^2**, and it is bounded as follows (**Fig. 1.1**, **Plate 1**):

- towards the west; by Høydalsfjorden; and a line from Steinhovden to Barlindbotn,
- towards the north; by the road section from Barlindbotn to Kalsvik with adjacent non-road exposures; by Eikefjorden; and by a line going north of and parallel to the road between Sunnarvik and Osa; including the road sections present from Sunnarvik to near Eikefjord church, with adjacent non-road exposures.
- towards the east; by the lake Vassetvatnet and a line continuing further southwards across Lyngskorfjellet to the Standal Fault.
- towards the south; by the Standal Fault, from Steindalen and westwards to Simavatn; and by a line going nortwestwards from Simavatn, continuing along the southern shore of the Langeneset penninsula.

Field work and data base

A total number of 297 days (~10 months) were spent on localities in the field, to cover the aims of the project. High and steep topographic relieff combined with exceptionally large amounts of heavy rain and fog (due to high mountains and short distance to the Ålfoten glacier!) reduced the effectiveness of field work in the area.

To maintain personal safety, caution must be executed when field work is conducted in the high grounds. Large parts of the Håsteinen Devonian Massif is bounded by steep mountain cliffs several hundred metres high. In case of fog, it is crusial to know that certain parts of the high grounds have only one, or a few, exit points that will lead safely to the lowland. These high grounds are briefly mentioned in the following.

The situation is particularly difficult in the southern parts of the massif. The mountain ridge going from the peak of Blåfjellnipa to Blåfjell (**Plate 1**) offers only one safe exit, which is southwards along the river located just southeast of the peak of Blåfjell. The mountain plateau further east that contains the peaks of Vestre and Østre Håsteinen have three exit options. i) Breiskåra, which is a narrow shelf in the steep mountain side below the peak of Vestre Håsteinen. This alternative is very hard to find in fog, and cannot be recomended. ii) A route across the Kvangagjelet syenite along a south trending pathway (called Øykeskåra, name not shown on

Plate 1) that is located just east of the river coming from lake Håsteinsvatnet, but this path can also be hard to find in a fog. The safest return will be iii) to go eastward from the peak of Østre Håsteinen and down to the valley Lelisdalen, where the river will lead eastwards to safe grounds.

Also in the northern parts, the Håsteinen massif is bounded by steep cliffs. From the ridge of Vikafjell, between the peaks of Mannen (in the west) and Vikanipa (in the east), any return routes down towards Osen in the central parts of the study area are normally safe, although numerous smaller local cliffs must be avoided. In the northward direction, only one safe exit option exists, located in the area just east of Mannen. (An exit option also exists downwards along the narrow cleft of "Galmannsskåra", but the cleft may be hard to find in foggy conditions and therefore is not safe). In the east, safe returns can be made eastwards from the peak of Vikanipa and in addition through the valley of Vassetbrekka. Another mountain plateau is present in the eastern part of the study area, containing the peaks of Leirvågfjellet, Nonsnova and Terigafjell. From this area, three safe return routes are present: i) through the valley of Vassetbrekka, ii) eastwards from a point **200 m** north of Nonsnova, where a steep valley goes down to the headland Lyngneset at lake Vassetvatnet, and iii) in the direction westwards to Osen.

The extensive field work has yielded a large data base. A selection of subjects has therefore been made, and the present thesis concentrates on the data related to the main elements of the geological history in the Håsteinen area.

1.2 ORGANISATION OF THE THESIS

Subsequent to this brieff introductory chapter (Ch. 1), the thesis is organised as follows:

Chapter 2 - Regional geology, provides an overview of the geology in the surrounding parts of western Norway. The chapter presents the three major units in the area, which are; (1) the basal gneisses of Precambrian origin (bottom of the tectonostratigraphy), (2) the Caledonian nappes, and (3) the Devonian sedimentary basins (top). Some models of regional significans that have been previously published in the literature, are discussed and challenged.

Chapter 3 - Eikefjord Group, EG, contains a description of (1) the lithologies of the group within the study area, (2) the structural development, (3) the metamorphic development, and in addition, (4) a description of the Sunnar Fault.

Chapter 4 - Høydalsfjorden Complex, HC, contains a description of (1) the lithologies, (2) the structural development, and (3) the metamorphic development.

Chapter 5 - Håsteinen Devonian Massif, HDM, gives a description of (1) the general sedimentology, (2) the primary contacts between the HDM and the substrate, (3) the substrate-inliers within the HDM, and (4) the structural and temperature/burial (metamorphic) history of the massif.

Chapter 6 - Summary, Discussion and Conclusions, (1) gives a summary of the main data of each of the three tectonostratigraphic units, (2) provides a summary of the chronology of geological events, (3) introduces a new model for the development of the Devonian massif, tested by numerical modelling, (4) discusses various aspects of the Devonian deformation in the study area, including development of the Nordfjord–Sogn Detachment Zone, and the effects of Devonian folding on the three units, and (5) concludes in the form of an integrated geological history of the whole area.

1.3 PRACTICLE INFORMATION

Maps

The mapping was carried out on the following 1:5000 "economical maps" ("Økonomisk Kartverk" photogram 1970) obtaind from "Fylkeskartkontoret i Sogn og Fjordane", Hermansverk:

Barlindbotn	AH	084-5-3	Vikarimma	AH	083-5-2	Svardal	AJ	083-5-3
Notaholmen	AH	084-5-4	Vasset	AJ	083-5-1	Gamlestølen	AH	082-5-1
Eikefjord	AJ	084-5-3	Høydal	AG	083-5-4	Håsteindalen	AH	082-5-2
Steinhovden	AG	083-5-2	Remma	AH	083-5-3	Steindalen	AJ	082-5-1
Pollen	AH	083-5-1	Osen	AH	083-5-4			

The economic 1:5000 field-maps form the basis for the structural maps in **Appendix A**.

The economic maps have no contours or other information in the areas situated above the **400 m** contour, and large parts of these maps were therefore supplied with 1:5000 enlargements of the following 1:50 000 maps of the "topographical main map series - M711" ("Topografisk hovedkartserie - M711, editions 1976 and 1994) with UTM coordinates, published by the Norwegian Mapping Authority ("Statens Kartverk"):

"Eikefjord", sheet 1118 II, grid zone 32V, 100km square KP + LP
"Naustdal", sheet 1218 III, grid zone 32V, 100km square LP

The **Road logs no**. **1**, **2** and **3** (**Appendix B**) from the western part of the study area are based on 1:1000 maps covering the road itself plus a narrow strip on each side. These maps were obtained at "Teknisk Etat" in Florø Kommune and were originally made for road construction.

References to localities on the enclosed maps are done by geographical names (see also Sect. 1.4) or by UTM coordinates. The UTM coordinates may be found on **Plate 1**, **Plate 3**, and on the 1:7000 structural maps in **Appendix A**.

The following aerial photographs have been used:

Scale 1: 40.000:			Scale 1:15.000:		
3400	804	D.13 – D.16	S.70	803	C.18 – C.34
3400	804	E.13 – E.17	S.70	803	D.17 – D.32
			S.70	803	E. 27 – E.41

Viewed together, the photographs with scale 1: 40.000 cover the whole study area. Also the photographs with scale 1:15.000 cover the whole study area, but with two local exceptions; a) the top plateau of the Håsteinen mountain, situated in the southernmost part of the study area, which escaped the aerial cameras due to high elevation (~1000 m) combined with large distance to the two nearest passing “photographing flight routes”, and b) the limited area to the north of Barlindbotn, which is outside the area covered of the photograph series listed.

Methods

Structural measurements were taken with a 360° Silva Type 15 T ("The Ranger") clinometer compass. No corrections have been made for the magnetic deviation (i.e. declination) from true north. At the time of measurement, the declination was **4.15 °W**, i.e. within the error limit of field measurements. Structural elements are plotted stereographically on equal area Schmidt stereo nets, lower hemisphere.

414 rock samples were collected in the field, and about **200** thin-sections have been investigated under polarizing microscope during the project. Interpretation of the metamorphic and structural development is based on direct field observations and thin-section microscopy of samples.

Terminology/definitions

In the present thesis, the expression "Håsteinen Devonian Massif" is applied to the ***western half*** of the massif (i.e the area west of lake Vassetvatnet) unless otherwise stated.

Structural data are given as follows: Planar structures are defined by their strike direction given by 0–180° and the amount of dip by 0–90° towards N/E/S/W, etc. (e.g. 120/60 SW means a surface with strike direction 120° and dip 60° towards southwest). For linear structures (fold axes and lineations) the whole circle from 0–360° are used for azimuth, and 0–90° for plunge, and the structure is plunging in the azimuth direction (e.g. 285/30 means a linear structure plunging 30° towards west-northwest.).

The different "generations" of structural elements and metamorphism are denoted as follows:

During the different generations of Deformation = D_n
the structures formed are: Foliation = S_n (S_0 = sedimentary bedding)
Lineation = L_n
and Folds = F_n
Axial planes are denoted = AP_n
The corresponding metamorphism is denoted = M_n
where "n" is the number of the generation.

It should be noted that the numbering of deformation phases relates exclusively to each tectonostratigraphic unit in the study area (which contains three such units: the Håsteinen Devonian Massif, the Høydalsfjorden Complex and the Eikefjord Group), and only reflects the structural history within this particular unit. Based on the structural interpretation of the present thesis, this means that e.g. *D_1-structures in one unit do not correspond to D_1-structures in another unit, and so on.* This is further discussed throughout the thesis.

Terminology of deformation structures are used mainly according to Ramsay (1967), Hobbs, Means and Williams (1976), Ramsay & Huber (1983, 1987) and Davis & Reynolds (1996). Interlimb angles of folds are classified as gentle (180–120°), open (120–70°), close (70–30°), tight (30–0°), and isoclinal (0°). Mylonites are classified according to Sibson (1977). Tectonites are given names that indicate the protolith when possible, i.e. "Meta-greywacke", "meta-psammite", "meta-gabbro", etc. These names are replaced by "fsp-qz-mica-schist", "greenschist", etc., only when the alteration is beyond recognition of the original rock. Magmatic rocks are named according to the classification of Streckeisen (1976). Shelley (1975), MacKenzie & Guilford (1980), MacKenzie et al. (1982) and Gribble & Hall (1985) have been used during the microscopical work. Terms used for the description of metamorphic textures are based on numerous books which will be referenced in the text. Lithological units (and to a certain degree structural elements) have been named according to the "Rules and recommendations for naming geological units in Norway" (Nystuen 1989).

1.4 GEOGRAPHICAL INFORMATION

Geographical names (Place names)

Geographical names used in the present thesis have been taken mainly from the 1:5000 economic maps, but also from the two 1:50 000 M711/UTM maps “Eikefjord” and “Naustdal”, listed above. In the following, some of the names will be given closer attention, due to changes in their use and orthography:

“Håsteinen”: As seen on **Fig. 1.1**, the mountain “Håsteinen” – which has given name to the Håsteinen Devonian massif **–** is located on the southern margin of the study area. The mountain forms a plateau with a western peak (**933 m**) and an eastern peak (**965 m**). The western peak is a steep and monumental landmark when seen from the west. Geologists have traditionally – and apparently in accordance with local practise – used the name “Håsteinen” for this western top, and the same name has also been used for the whole plateau including both peaks. When necessary, the western top has been called “Ytre” (Outer) or “Vestre” (Western) Håsteinen , and the eastern one “Indre” (Inner) or “Østre” (Eastern) Håsteinen. Complying with this, the 1976 edition of the 1:50 000 M711 map “Eikefjord” has the name “Håstein” for the western peak, and no separate name for the eastern peak. On the 1994 edition of this map, however, the name of the western peak has been changed from “Håstein” to “Høydalsnipa”, whilst the name “Håsteinen” is now in stead placed on the eastern peak. In the present thesis, the traditional naming will be maintained (**Fig. 1.1**), using “Vestre Håsteinen” for the western peak and “Østre Håsteinen” for the eastern one.

Several other names in the study area have also been changed on the 1994 “Eikefjord” M711 map compared to the 1976 edition, and also compared to the economic maps. In the present work, some of the traditional pre-1994 names will be sustained, as in the above case for “Håsteinen”. The preferred names are listed in the following table (in bold):

Name, and geographical position	1976 map edition	1994 map edition	Economic maps	This work
Håsteinen, at southern margin of the study area	- Western peak: **Håstein** - Eastern peak: No name	- Western peak: <u>Høydalsnipa</u> - Eastern. peak: Håsteinen		**Vestre Håsteinen** **Østre Håsteinen**
Nonsnipa, S of Osen in central area	**Nonsnipa**	No<u>sa</u>nipa	**Nonsnipa**	**Nonsnipa**
Nonsnova, W of lake Vassetvatnet	**Nonsnova**	No<u>sa</u>nova		**Nonsnova**
Lelisdalen, SE margin of study area	**Lelisdalen**	L<u>i</u>lisdalen		**Lelisdalen**
Vikafjell, on N mountain plateau	**Vikafjell**	Vikafjell<u>et</u>		**Vikafjell**
Straumsnes, at inlet Osstrupen fjord	**Straumsnes**	Straumsnes<u>et</u>	**Straumsnes**	**Straumsnes**
Teigafjell, W of lake Vassetvatnet	**Teigafjell**	Teigafjell<u>et</u>		**Teigafjell**
Håsteindalen, SE marg. of stud.area	**Håsteindalen**	Håstein<u>s</u>dalen	**Håsteindalen**	**Håsteindalen**
Vasset, W of N end of lake Vassetv.	Vassete	**Vasset_**	**Vasset**	**Vasset**
Vassetdalen, W of lake Vassetvatn.	-----	Vasset<u>e</u>dalen	**Vassetdalen**	**Vassetdalen**
Vassetvatnet, E margin of stud. area	Vassetevatnet	Vassetevatnet	**Vassetvatnet**	**Vassetvatnet**

Table showing geographical (place) names that have been subjected to recent changes in othography and use. Names in bold are used in the present thesis. (Underlined parts of 1994-names indicate deviations from the 1976-names).

Geography

The area investigated in the present thesis is located in Flora Kommune (municipality) in the region of Sunnfjord in Sogn og Fjordane Fylke (county), Western Norway. The village of Eikefjord at the northernmost boundary of the study area (**Fig. 1.1**) is the nearest place which offers facilities like fuel station, food store (with postal services), etc. Local overnight accomodation may be offered by “Sunnfjord Hyttegrend & Camping” (Sunnfjord cabins and camping) located at Storebru, 4.5 km (road distance) to the east-southeast of Eikefjord. The nearest town is Florø, located about 25 km (road distance) to the west of Eikefjord. Florø offers a wide range of overnight accomodations.

In the study area, farms and other buildings are situated mainly along the coast, or, when inland from the coast, in the lowland in the valleys. Access to the study area is provided by three roads that transect the area, termed fylkesvei 541, 542 and 543.

In the west, along the fjords of Eikefjorden and Høydalsfjorden, the road termed fylkesvei 542 provides fairly fresh road sections through all the three major rock units. The road crosses the Osstrupen fjord on a concrete bridge, and road sections in this area offers good exposures of the Devonian-substrate contact.

From a road junction at Sunnarviken, a narrower road termed fylkesvei 541 extends eastwards to Osa at lake Vassetvatnet. Midway between Sunnarviken and Osa, a narrow road goes southwards to Vassethola. From Osa, the 541 road passes Vasset on its way southwards along the western shore of lake Vassetvatnet which defines the eastern margin of the study area. The road-sections to the south of Vasset are older and, hence, more lichen- and moss-covered. In Svardal this road splits into two; one road goes westwards to Osen and the other eastwards into the valley of Steindalen.

At Pollen in the NW part of the study area, the road termed fylkesvei 543 takes off norhtwestwards from fylkesvei 542. After 500 m, fylkesvei 543 splits into two: one road goes to Barlindbotn and Kalsvik, and the other one towards Steinhovden, continuing to Sandvika outside the study area. This road offers fairly fresh road sections.

Locally in the study area, tractor roads takes off from the main roads. For example, a tractor road is present i) at Store Høydal where it goes eastwards up into the valley; ii) at Sunnarviken where it goes southwards into the hillside below Vikanipa; and iii) south of Vasset where it goes westwards into the Vassetdalen valley.

From the main road Florø–Førde (riksvei 5), at a point about 7 km east of Eikefjord, a narrow road goes southwards into Agledal. This road, and the road to Steindalen, provide access to the eastern half of the Håsteinen Devonian Massif.

Geomorphology

The study area is located in a fjord landscape typical of the county of Sogn og Fjordane. The narrow, east–west directed fjords of Eikefjorden and Høydalsfjorden, where the latter continues in the form of Osstrupen, reach into the study area from the west. E-W trending valleys represent continuations of the fjords. The fjords and valleys were formed by glacial activity during Quaternary times.

The geomorphology of the study-area is closely related to the rocks' abilities to resist weathering and erosion, and these abilities depends on mineral composition and structural characteristics. Geomorphological/topographical features are shown in **Fig. 1.2**, **Fig. 1.3** and **Fig. 1.4**. Rocks with a high content of quartz and/or feldspar tend to be most resistant to weathering and usually form topographic highs. The low and wide valley of Eikefjord is formed in the mylonitized gneissic Precambrian rocks of the Eikefjord Group, between the Eikefjord and Sunnar Faults. It is possible that the penetrative low grade mylonite fabric has weakened the rock. Around the elevated ridges of the Håsteinen Devonian Massif, the relatively low hills and open valleys consist of the Lower Palaeozoic meta-greywackes and meta-semipelites (both intruded by meta-gabbro bodies) and meta-psammites, of the Høydalsfjorden Complex (**Fig. 1.2**). Larger meta-psammitic areas tend to form topographic highs. On the hillside rising upwards from the fjord of Eikefjorden in the direction towards eastern part of Vikafjell, two WNW-ESE- trending vertical escarpments, with a height of minimum **10–**

20 m and a length of up to **2 km**, are clearly seen (**Fig. 1.2**). The escarpments can only be transversed a few places, and are possibly related to mylonitic horizons in the hillside. The conglomerates of the Håsteinen Devonian Massif, with pebbles of meta-psammites and vaguely deformed syenite and gabbro, form the most spectacular topographic relieff in the area, as the massif rises steeply from the fjords or the surrounding low land. The Vikafjell area (**Fig. 1.2**), the Leirvågfjellet–Teigafjell area and the Blåfjellnipa–Håsteinen–Fjellsenden area (**Fig. 1.3**) form ridges and plateaus with elevations ranging from about **550** to about **950 m**. As mentioned above, the highest mountain in the study area bears the name Håsteinen, with the two peaks Vestre Håsteinen (**932 m**) (**Fig. 1.3**) and Østre Håsteinen (**965 m**). With its position southeast of the fjord Osstrupen, the very Håsteinen mountain rises steeply directly from the fjord. Osstrupen, and the valleys containing the lakes of Svardalsvatnet (**20 m.a.s.l**.) and Vassetvatnet (**27 m.a.s.l**.), make deep cuts through the Devonian massif (**Fig. 1.4**). The Devonian rocks form steep to vertical mountain cliffs which are usually **300–500 m** (locally **>700 m**) high along the sides of these valleys and along the outer margins of the Devonian rocks (**Fig. 1.2** and **1.4**).

Degree of exposure

The degree of exposure in the study area varies considerably, depending on lithology and altitude. The Eikefjord Group rocks are usually covered by forests of pine trees with scattered outcrops, or by farmland without exposures. The rocks of the Høydalsfjorden Complex are generally extensively covered by pine or birch forests below the timber line (~ 400 m), and the degree of exposure is locally low. An exception from this is found in the areas with meta-psammites and meta-greywackes, which are better exposed. Above the timber line, the degree of exposure is generally higher. The rocks of the Håsteinen Devonian Massif are well exposed, reaching up to near 100% exposure on the elevated plateaus and ridge areas.

Quaternary deposits

In the study-area there are several areas with significant Quaternary cover. At Svardal and at Osen, remnants of late Weichselian glacifluvial marine deltas, now forming flat terraces, are present, together with deposits caused by later fluvial reworking of these terraces. The Quaternary deposits in these two areas cover the contact between the Devonian rock and it’s substrate. The most extensive Quaternary cover in the study area are present in the valley going from the sea shore at Store Høydal and further eastwards up the valley. In the lowest, westernmost, part of the valley, terraces of glacifluvial marine delta deposits are present, and further up the valley, diamicton deposits take over. These two types of deposits completely cover the valley floor, thereby also covering the contact between the Devonian rocks and the substrate. Towards the east, the diamicton deposits are gradually replaced by the large talus cones at the base of the Vestre Håsteinen mountain peak. At the base of this peak, the scree deposits cover the contacts between the Håsteinen Devonian Massif, the Kvangagjelet Syenite and the Høydalsfjorden Complex. From the Høydalsstølen area and eastwards, the diamicton is replaced by extensive bog fields, and these continue towards the lake Høydalsvatnet. Some exposures are present in the bog field, particularely along the river. Major talus cones are also present at several other locations along the base of the Devonian cliffs (**Plate 1**, and **Appendix A**), e.g. to the north of Vikafjell, below the cliffs at Litleteigen, in the steep slopes just to the west of the lake Vassetvatnet, at the southern part of the Gravaneset area, and other places.

1.5 PREVIOUS WORK

The present section gives an overview of the publications dealing with the Håsteinen area. The works are listed chronologically to show the historical development of the geological knowledge of the area. The list contains works dealing specifically with the geology of the Håsteinen area, together with more regional works containing limited data from the Håsteinen area.

(1) Irgens & Hiortdahl (1864) presented the very first geological map showing the location of the Håsteinen Devonian Massif, as well as the other Devonian massifs. Their account of the geology of the area is very brief and general, and the parts of the Håsteinen Devonian Massif located to the east of the lake Vassetvatnet were not recorded on their map. They were the first to suggest a Devonian age for the sediments.

(2) Kjerulf (1879) gives an overview of the geology of south Norway, and mentions the Devonian basin of Håsteinen.

(3) Helland (1880) briefly described some of the rocks in the thesis area and suggested a structural configuration. He was the first to note the synclinal form of the Devonian basins and suggested a fold pattern along the coast between Sognefjorden and Nordfjord, where the synclines containing Devonian sediments were separated by corresponding anticlines between the basins.

(4) Reusch (1881) investigated the West Norwegian Devonian massifs and substrata. From the Håsteinen area he gave very brief petrographic descriptions of the rocks at Eikefjord, Sunnarvik, Vassetbrekka, Osen and Svardal. He particularly studied the Devonian conglomerates at Osen at the end of the fjord Osstrupen. He presented the best map of the Devonian basins so far, showing rock boundaries, strike-dip symbols and areal distribution of conglomerates and sandstones.

(5) Kolderup (1925) presented the first separate investigation of the Håsteinen Devonian Massif, including the first separate geological map. Although the investigation was of reconnaissance type, fairly detailed petrographic rock descriptions are given for both the Devonian sediments and the substrate. The primary unconformity in the west was noted.

(6) Kolderup (1928) gave fairly detailed petrographic descriptions of the rocks situated in the substrate below the Devonian rocks. Field descriptions are supported by accounts of the mineralogy based on thin-section studies. He described the meta-anorthosites in the Eikefjord area, as well as the Lower Palaeozoic rocks located between the fjords of Eikefjorden and Brufjorden.

(7) Bryhni (1958) essentially dealt with the area to the north of the Sunnar Fault, but also included observations from the Håsteinen Devonian Massif and the Lower Palaeozoic rocks below. He gave detailed

petrographic descriptions of the Precambrian rocks in the Eikefjord area, and also of the Lower Palaeozoic gabbros in the area between Barlindbotn and Høydalsfjorden. Descriptions are given of the Lower Palaeozoic rocks to the west of the lake Vassetvatnet, and of Devonian rocks and their substrate to the east of the lake Vassetvatnet.

(8) Kolderup (1960) gave a brieff account of the Håsteinen Devonian Massif and its substrate. In this work, which is an excursion guide, he emphasised the synclinal structure of the four main Devonian basins in Western Norway. He further reported that the Håsteinen Devonian Massif was situated with a primary unconformity on its substrate, and that later tectonic movements locally have disturbed this unconformity. Just to the west of the lake Vassetvatnet, between the headlands of Strupeneset and Stigen, he noted that the substrate rocks "rise" high up "into" the Devonian rocks. He interpreted this feature as an anticline structure related to a phase of post-Devonian N-S compression which also generally folded the Devonian massifs of western Norway. He also noted the low degree of metamorphism in the Devonian sediments.

(9) Bryhni (1962) carried out structural investigations of the area just to the north of Eikefjord, i.e an area directly bordering the area investigated in the present thesis. In this work, Bryhni reports for the first time on deformation phases and on related generations and orientations of foliations, lineations, fold axes, axial planes, etc., beeing the first to apply the method associated with $D_X/S_X/L_X/F_X/AP_X$ terminology and stereographic techniques, combined with microscopic fabric studies — to rocks in these parts of Sunnfjord.

(10) Bryhni (1964a) discussed the consequences of the presence of mylonites in the area between the Håsteinen and Hornelen Devonian Massifs. He also reported very briefly on the metamorphic and structural state of the Devonian conglomerates in the road sections at the headlands of Stigen and Strupeneset at the lake Vassetvatnet.

(11) Bryhni (1966) presented the first attempt to subdivide the gneisses situated in the area between the coast of Sogn og Fjordane and the Jotun Nappe Complex. He introdused the name "Jostedalen Complex" for the largely migmatitic rocks in the eastern part of the gneiss area, and the name "Fjordane Complex" for the gneisses and meta-supracrustals to the west. The anorthosites and other gneisses situated in the northern part of the area studied in the present thesis are part of his Fjordane Complex.

(12) Kildal (1970) presented main features of the geology of the Håsteinen area on the geological 1:250.000 map Måløy, which covers the area between Sognefjorden and Måløy.

(13) Osland (1970) studied the Sandviken meta-psammite located 5 km to the northwest of the Håsteinen Devonian Massif. In this work he demonstrates for the first time that the rock is actually a meta-psammite and not a granite as reported in all earlier publications. He also collected samples from the meta-psammite at Osen — which forms a substrate inlier within the Håsteinen Devonian Massif — and used them for petrographic and geochemical comparison. In addition he presented a geological map at a scale of 1:100.000 covering the large area located between the Standal Fault and the Hornelen Devonian Massif.

(14) In Bryhni et al. (1981), Bryhni presented a further twofold subdivision of the Fjordane Complex (of Bryhni (1966), see above) into the Eikefjord Group and the Lykkjebø Group. In this excursion

guide he reported that the Eikefjord Group consists of meta-anorthosites, syenitic to monzonitic biotite-epidote-gneisses, meta-gabbros, amphibolites and meta-ultramafites. He also stated that the metamorphism of e.g. syenites/monzonites have been retrogressive from granulite facies via epidote-amphibolite facies to greenschist facies. The Lykkjebø Group was reported to consist of feldspathic quartzite and feldspathic mica-schist, both with some dolomite marble. Rocks of the Lykkjebø Group were reported to be locally interlayered with rocks of the Eikefjord Group. Bryhni also presented a geological 1:100.000 map covering the Eikefjord Group in the area north- and eastwards from the lake Vassetvatnet. The area studied in the present thesis contains rocks of the Eikefjord Group in the northernmost part. It may also be noted that further south, at the road-tunnel entrance through the headland of Strupeneset at the southern part of lake Vassetvatnet, the contact between the Håsteinen Devonian Massif and the substrate was briefly described as "phyllonitized".

(15) Abdel-Monem & Bryhni (1978) reported a radiometric Rb/Sr whole rock age for the meta-igneous rocks of the Eikefjord Group, based on samples taken from the Eikefjord area and the Gloppen area. The dating yielded an age of 1511 +/- 64 Ma which was interpreted as minimum age of intrusion.

(16) Torsvik et al. (1987) made an investigation of the palaeomagnetism of the Håsteinen Devonian Massif and its substrate, and also of tectonomagnetic fabrics and related susceptibility ellipsoids which formed during the deformation of the Devonian rocks. A geological map from the wersternmost parts of Håsteinen was presented.

(17) Seranne (1988) presented a regional investigation of the formation and deformation of all the Devonian massifs in western Norway as well as the detachment zone below the massifs. The work also includes some data from the Håsteinen area.

(18) Vetti (1988 - abstract) presented, for the first time, a structural analysis of the Håsteinen Devonian Massif, showing that the basin is strongly folded about a fold axis plunging 50–60˚ towards ESE, and that the fold has a northern limb oriented with a strike/dip of about 070/60 SSE, and a southern limb oriented about 160/60 ENE. For the first time the bedding-normal stratigraphical thickness could be estimated to c. 6 km. New data were also presented on sedimentology, inlier bodies, contact relations, etc. In addition, new data were also presented from the subjacent Caledonian ophiolithic rocks and Precambrian gneisses.

(19) Chauvet & Seranne (1989) made a regional study of the microtectonic evidence for Devonian extensional top-to-west shearing in southwest Norway. They presented a structural map of the Sognefjord-Nordfjord region, also containing a few sense-of-shear measurements from the detachment mylonites of the Håsteinen area.

(20) Seguret et al (1989) gave a report on the characteristics of, and governing principles for the formation of, the new "collapse basins" of western Norway. They presented regional maps of the Sognefjord-Nordfjord area, also containing some data from the Håsteinen area, notably a few shear-sense measurements from the detachment zone, and a few extension direction indications from the Devonian rocks.

(21) Seranne et al. (1989) discussed the regional tectonics of the Devonian "collapse" basins of Western Norway. They presented structural maps of the Sognefjord-Nordfjord region which included a few

measurements of bedding orientation in the Håsteinen Devonian Massif; directions of shear sense and lineations in the substrate; extension direction in the Devonian rocks and the substrate; dip of the detachment along the southern margin of the Devonian rocks; and direction and sense of movement of the upper plate to this detachment.

(22) Vetti (1989 - abstract) presented much of the same data as Vetti (1988). In addition to the overall structural geometry of the Håsteinen basin, the sedimentology was a subject. This included distribution of conglomerate and sandstone facies. Three inlier bodies of substrate rocks were presented. Attetion was also drawn to the axial planar cleavage related to small-scale folding in the westernmost sandstones.

(23) Andersen & Jamtveit (1990) studied the structural development of the Nordfjord–Sogn Detachment below the Kvamshesten Devonian Massif and also at the southern margin of the Håsteinen Devonian Massif. The paper contains a block diagram giving a three dimensional model of their view on the relationship between the Nordfjord–Sogn Detachment Zone and the Håsteinen and Kvamshesten Devonian Massifs.

(24) Bryhni & Lutro (1991a) presented the main features of the geology of the western part of the Håsteinen area on the geological 1:50.000 map EIKEFJORD (*second* preliminary edition, Norges geologiske undersøkelse). The map covers the area ranging from the Håsteinen Devonian Massif into the Hornelen Devonian Massif, and contains a profile crossing the Håsteinen Devonian Massif. Note: The map is no longer available at NGU (Norwegian Geological Survey), because it has been replaced by the revised *third* edition: Bryhni & Lutro (2000a), see below.

(25) Bryhni & Lutro (1991b) presented features of the geology of the eastern part of the Håsteinen area on the geological 1:50.000 map NAUSTDAL (*second* preliminary edition, Norges geologiske undersøkelse). The map covers the area from the Håsteinen Devonian Massif to the Hornelen Devonian Massif. Note: The map is no longer available at NGU (Norwegian Geological Survey), because it has been replaced by the revised *third* edition: Bryhni & Lutro (2000b), see below.

(26) Lutro (1991) presented the main geology of the municipality Flora Kommune on the geological 1:75.000 map FLORA (preliminary edition, Norges geologiske undersøkelse), also covering the Håsteinen area.

(27) Chauvet & Dallmeyer (1992) performed a regional geochronological study, obtaining $^{40}Ar/^{39}Ar$ mineral dates on samples from the detachment mylonites presently situated around (and below) the Devonian basins of western Norway. (Also samples from the adjacent Western Gneiss Complex were dated). Two of the samples were taken from the area between the Hornelen and Håsteinen Devonian Massifs, an area beeing part of the mylonite zone and comprising rocks of the Lykkjebø and Eikefjord Groups. The ages ages obtained were 399.2 +/- 0.7 Ma and 402.9 +/- 1.1 Ma.

(28) Fossen (1992b) made a regional study on Caledonian contractional (top-E) and Devonian extensional (top-W) shear sense indicators in central and western parts of south Norway. The paper contains maps including a few measurements also from the Håsteinen area (compiled from Seguret et al. 1989). The

Devonian large-scale extension of western Norway were devided into two different modes: Mode I, westward movement of the orogenic wedge on the basal decollement, and Mode II, development of low-angle westward dipping detachment zones (e.g. Nordfjord–Sogn Detachment Zone) cutting the orogenic wedge and the basal decollement. It was argued that the Devonian extensional movements occurred as a result of divergent plate motion between Baltica and Laurentia.

(29) Andersen’s (1993) “Discussion” of Fossen’s 1992b-paper, disagreed with several of Fossen’s (1992b) conclusions. Andersen (1993) stated that: a) Devonian extension in the hinterland was caused by internal body forces (gravity driven) generated within the thickened lithosphere, and not divergent plate motions between Baltica and Laurentia, as suggested by Fossen (1992b). b) Westward extension in the *hinterland* was contemporaneous with deposition and eastward thrusting of the Ringerike sandstone at the *foreland*, i.e., opposing Fossen’s (1992) view that the *whole* orogenic wedge was moving westward from the time the hinterland extension commenced.

(30) Fossen’s (1993) “Reply” to Andersen’s (1993) “Discussion”, stated the following: a) Gravity driven hinterland extension, and foreland thrusting of the Ringerike sandstone may well have occurred at an early stage, but the reversal of the shear sense along the decollement zone clearly postdates this deformation, as does the extension along the NSD. b) The westward movement of the orogenic wedge into the central part of the orogen is not possible unless “space” is simultaneously beeing provided by some sort of divergent plate motion. c) Devonian basins (e.g at the Møre Tøndelag Fault Zone) may have formed due to local extension within the orogenic plate, *prior to* westward movement of the whole orogenic wedge.

(31) Chauvet & Seranne (1994) conducted a regional study in the Sognefjorden– Nordfjord area, focusing on the late-orogenic Devonian folding which has effected the upper plate (including the Devonian basins), the Nordfjord–Sogn Detachment Zone and the lower plate. On a regional map they presented data from the Håsteinen area on contractional (Caledonian) lineations, extensional lineations and fold orientations. They concluded that the folds formed during Devonian ductile extension and Old Red Sandstone basin formation.

(32) Wilks & Cuthbert (1994) studied the top-to-west-sheared rocks of the Nordfjord–Sogn Detachment Zone around the Hornelen Devonian Basin. They recorded shear sense and lineation/ foliation orientations from the mylonitic detachment zone, notably also from the area between Håsteinen and Hornelen. This area contains rocks of the Eikefjord and Lykkjebø Groups, which also continue into the study area of the present theses.

(33) Vetti (1996 - abstract) presented data from the Håsteinen Devonian Massif, and in addition data from the subjacent ophiolite-related rocks (now named the Høydalsfjorden Complex), and the Precambrian mylonitic rocks (Eikefjord Group), part of the Nordfjord–Sogn Detachment Zone. Fokus, however, was on the formation of the Håsteinen Devonian basin itself and the deposition of the ORS sediments. For the first time the extraordinary geometrical situation in the Håsteinen basin was presented: the high angle between the steeply dipping Devonian *beds* and the subhorizontal *unconformable contact* (envelope surface) towards the underlying rocks of the Høydalsfjorden Complex. A new model for formation of Devonian basins was presented to explain this geometry, **“The ramp syncline rotation model”:** the Håsteinen basin must have formed on top of the

Upper Plate in a ramp depression (ramp syncline), which developed above a frontal ramp in the underlying Nordfjord–Sogn Detachment Zone.

(34) Vetti (1997a - abstract) presented some of the same data and model as appared in Vetti (1996). However, the name of the model was now simplified to **“The ramp model”**. A frontal ramp responsible for the Håsteinen basin must have been located east of the present position of the Håsteinen Devonian Massif and the subjacent Høydalsfjorden Complex. The ramp presumably died out towards the north, i.e towards the Hornelen basin.

(35) Vetti (1997b - excursion guide) gave an overview of the geology of the Håsteinen and Hornelen areas. From the Håsteinen Devonian Massif, data were given on the sedimentology, vein intrusions, contact relationships, and inliers of substrate rocks. Fokus, however, was on the deformation of the massif, and data were given on the folded beds (the Osstrupen syncline), minor parasitic folds at Gravaneset, effects of Devonian deformation on the substrate rocks, stratigraphic thickness, metamorphism, and faults. Two sketch maps showed the general geology of the Håsteinen area, and the axial trace and bedding/limb orientations in the Osstrupen syncline of the Håsteinen Massif. A profile sketch (W-E) showed the principle used to calculate the bedding-normal stratigraphic thickness of 5,8 km.

(36) Andersen (1998) gave an overview of the extensional tectonics in the Caledonides of southern Norway. The discussions, as well maps and profiles, have relevans to the development of the Håsteinen basin.

(37) Fossen & Dunlap (1998) used $^{40}Ar/^{39}Ar$ on muscovite and biotite to date tectonic movements in the decollement zone below the Caledonian nappes. Top-to-the SE fabrics, interpreted as contractional deformation/ thrusting, occurred in the time interval **415-408 Ma**, and top-to-the-NW fabrics, interpreted as extensional deformation/hinterland-directed nappe translation (Mode I - extension), occurred in the time interval **402-394 Ma**. This detachment zone was later cut by the steeper-dipping Nordfjord–Sogn Detatchment zone (Mode II - extension), which is present below the Håsteinen Devonian Massif.

(38) Krabbendam & Dewey (1998) discussed whether the exhumation of UHP rocks could be explained by means of a constrictional strain model for transtension in the Western Gneiss Region. The publication mentions the Håsteinen area, and also presents data from the Standal fault just west-southwest of this area.

(39) Braathen (1999) discussed the kinematics of post-Caledonian polyphse brittle faulting in the Sunnfjord region. In addition to the fault data, the study presented data from the easternmost end of the Grøndalen syncline, which is defined by folded bedding of the Hornelen Devonian Massif, and located along the southern margin of the Massif. The geometry of this syncline is quite similar to the Osstrupen syncline of the Håsteinen Devonian Massif, and a comparison of the two is made in the present theses.

(40) Eide et al. (1999) presented $^{40}Ar/^{39}Ar$ results from the areas west and south of the Kvamshesten Devonian basin, with samples from the Høyvik group, the Nordfjord–Sogn detachment zone, and the Western Gneiss region. The first and last of these sample localities revealed a rapid cooling event of **15**

°C/m.y. in Late Devonian–Early Carboniferous time (**360–340 Ma**), which was interpreted as an episode of unroofing, possibly caused by increased topography and erosion (accompanied by, or post-dating, folding around E-W axes and low-grade metamorphism) resulting from regional, thermal underplating and regional, transcurrent tectonics. It was suggested that the cooling episode and associated deformation /metamorphism would fit well with the previously proposed "Solundian Orogeny" (and Svalbardian Orogeny).

(41) Bryhni & Lutro (2000a) presented the main features of the geology of the western part of the Håsteinen area on the geological 1:50.000 map EIKEFJORD (*third* preliminary edition, Norges geologiske undersøkelse). The map covers the area ranging from the Håsteinen Devonian Massif into the Hornelen Devonian Massif, and contains a profile crossing the Håsteinen Devonian Massif. Revisions made since the *second* edition, which came in 1991, are: the name "Naustdal Group", containing Precambrian rocks of the Jostedal Complex, has been changed to "Basement, mainly Precambrian rocks". The name "Sunnar Fault" has been changed to "Sunnarvik Fault".

(42) Bryhni & Lutro (2000b) presented features of the geology of the eastern part of the Håsteinen area on the geological 1:50.000 map NAUSTDAL (*third* preliminary edition, Norges geologiske undersøkelse). The map covers the area from the Håsteinen Devonian Massif to the Hornelen Devonian Massif. Revisions made since the *second* edition, which came in 1991, are: the name "Jostedal Complex", containing Precambrian rocks, has been changed to "Basement, mainly Precambrian rocks". The name "Sunnar Fault" has been changed to "Sunnarvik Fault".

(43) Fossen (2000) discussed whether the extensional tectonics (from Mode I and onwards) in the Caledonides were synorogenic or postorogenic, and concluded that the extension was postorogenic, i.e. caused by divergent plate movements. The Håsteinen Devonian Massif is located within the area discussed in the paper.

(44) Engvik & Andersen (2000) studied the Caledonian eclogite and amphibolite facies deformation fabrics at Vårdalsneset (Western Gneiss Region, WGR) south of the Kvamshesten Devonan basin. Calculations of the T/P condidtons that prevailed during rock formation, showed that the eclogites formed at **T = 680 ± 20 °C / P = 16 ± 2 kbar**, and **T = 690 ± 20 °C / P = 15 ± 1.5 kba**r, and that the amphibolite facies formed at **T = 564 ± 44 °C / P < 10.3–8.1 kbar**. The authors suggested that these T/P estimates be relevant to larger parts of the Sunnfjord region, i.e the area incorporating the WGR area south and east of the Håsteinen Devonian Massif.

(45) Osmundsen & Andersen (2001) interpreted the Middle Devonian basins of western Norway as a sedimentary response to large-scale transpressional tectonics, where the NW-ward directed extension in western Norway was rotated to a W-ward extension due to SW-ward directed sinistral strike-slip movements along the Møre–Trøndelag Fault Zone. The paper presents data from the Håsteinen Devonian basin, obtained from Vetti (1988, 1996, 1997) and Vetti & Milnes (1997).

(46) Braathen et al. (2002) gave an overview of the orogen-parallel extension of the Caledonides in northern Central Norway, but also included an overview of the orogen-orthogonal extension in Western Norway, which was responsible for the formation of Håsteinen and the other Devonian deposits there. With

respect to Western Norway, the authors convey the commonly held view that the three prosesses of a) extension, b) Old Red Sandstone basin formation and c) ORS basin folding, were apparently synchronous, and caused by for example crustal-scale constrictional strain, or by transpression/transtension related to movements along the Møre–Trøndelag Fault Complex, MTFC.

(47) Osmundsen et al. (2003) studied the Devonian Nesna Shear Zone and adjacent gneiss-cored culminations in the North–Central Norwegian Caledonides, but also made a review of, and comparison with, the general Devonian extensional development and related basins, in Western Norway. The WSW-directed extensional movements in the Nesna shear zone commenced around **398 Ma** (Eide et al. 2002). In western Norway, the onset of W-directed extension occurred at **402 Ma** according to Fossen & Dunlap (1998b), and at **395 Ma** according to Terry et al. (2000). This shows that extension in Central and Western Norway started fairly simultaneously, although movements were in different directions.

(48) Young et al. (2007) studied the prograde amphobolite facies to ultrahigh-pressure eclogite (UHP) facies transition in the inner Nordfjord region, and the implications for the exhumation of the eclogitic Western Gneiss Region (WGR). Structural, thermobarometric and radiometric data were obtained along the fjord Nordfjord just north of the Hornelen Devonian Massif, as well as in the area stretching eastward from the Hornelen massif, across the Sandane (Gloppen) region. The northern- and easternmost part of the study area contain rocks of the WGR. *Eastward* from the massif, the strongly sheared rocks of the Nordfjord–Sogn Detachment Zone (NSDZ) were studied, and the local segment of this zone was named the Sandane Shear Zone.The detachment zone was found to have developed within rocks of the Lower and Middle Allochthons, i.e. in the Eikefjord and Lykkjebø Groups (as well as in the upper part of the subjacent WGR).

Southwestward from the Sandane (Gloppen) region, i.e. *outside the area investigated by the authors*, the Eikefjord and Lykkjebø groups continue for **40 km** all the way to the Eikefjord area, and also another **20 km** westwards to the town of Florø. Along this entire distance, the two groups continue to constitute the NSDZ. In the Eikefjord region, the sheared detachment rocks of the Eikefjord and Lykkjebø Groups enter into the study area of the *present thesis.*

Young et al. (2007) performed radiometric dating (U/Pb, zircon) on an ultra-high pressure (UHP) eclogite sampled from a locality just north of lake Hornindalsvatn, yielding an age of of **405 +/- 2 Ma** for peak eclogite metamorphism. Thermobarometric analyses showed that, over a lateral distance of **20 km** from the Sandane (Gloppen) area and northwards, a continuous transition was found to exist from the amphibolite facies **(~1.5 Gpa/600–700 °C**) crust in the south, through quartz eclogite **(~2.4 Gpa/600 °C**), and into UHP eclogitic (~**2.7–3.3 GPa/700°C**) crust in the north. Based on this continuity, the authors concluded that the Norwegian UHP province remained attached to lower-pressure crust during exhumation, and that a large area of the Western Gneiss Region was exhumed as a relatively coherent body.

(49) Marques et al. (2007) applied ’inclusion behaviour models’ to the Nordfjord–Sogn Detachment Zone system, to assess vorticity, strain, narture of rigid inclusion/matrix interface and confinement. They investigated three localities within the shear zone: **i)** Gjervika at the island Atløy, located west of the Kvamshesten Devonian Massif; **ii)** Sandane to the east of the Hornelen Devonian Massif, and **iii)** Biskjelneset,

near Verpeneset north of the northwestern corner of the Hornelen Devonian deposits. The cross section accompanying the map figure transects the Håsteinen area, i.e. the study area of the present thesis, and the results of Marques et al. (2007) may apply to the NSDZ in general. Their main conclusions were: **1)** At site i), Gjervika, which is positioned at a structurally high level within the NSDZ, the Shape Preferred Orientations (SPO) were explained by *simple shear* associated with a slipping inclusion/matrix interface. **2)** At Site ii), Sandane, which is located at deeper levels of the NSDZ, SPO was interpreted to have been produced by *simple shear* associated with a significant amount of *shortening* across the shear zone, acting upon rigid inclusions in slipping contact with the enclosing matrix. **3)** At site iii), Biskjelneset, which was located structurally deepest of the three localities, the authors suggested that the observed back rotated boudins had formed in confined flow associated with a considerable amount of shortening across the shear zone. **4)** The observed tails of porphyroclasts were interpreted to indicate a minimum (at least local) strain of **γ ~20**. **5)** the NSDZ was reported to show evidence of strain partitioning: rocks could vary from protomylonites to ultramylonoites, and simple shear and pure shear components were heterogeneously distributed. **6)** Therefore, the authors concluded that "flow in the NSDZ was very heterogeneous both at the kilometre and metre scale". Furthermore, the data were interpreted to suggest that "the amount of shortening across the shear plane throughout the NSDZ increases with depth, and the flattening component contributes to exhumation of the eclogite facies rocks in its footwall".

(50) Johnston et al. (2007a) gave a review of structural, petrological (P/T) and geochronological data within and around the entire Nordfjord–Sogn Detachment Zone (NSDZ) and evaluated the different models for exhumation of ultrahigh-pressure (UHP) rocks in western Norway, particularly focusing on the role of the NSDZ. A roughly north–south trending cross-section was compiled from existing data in the area between the Solund region (outer Sogn) in the south to the Sørøyane islands (near Ålesund town) in the north, a distance of **150 km.** The section crosses the four Devonian massifs of western Norway and shows structural features and subjacent geological units. Hense, the section also includes the Håsteinen Devonian massif and subjacent units, which are discussed in the *present thesis*. North of the Hornelen Devonian Massif, the cross section shows how the UHP rocks of the WGR appear at the subaerial surface. The authors suggested a three stage model for the exhumation of the eclogitic Western Gneiss Region (WGR): **i)** rapid exhumation of a HP/UHP body through the mantle to the base of the crust; **ii)** exhumation from the lower and middle crust along an initially broad ductile shear zone, the NSDZ, which formed during stage two; and **iii)** exhumation along more descrete brittle–ductile faults that formed during stage three.

(51) Hacker (2007) presented a review of the radiometric and petrologic (P/T) data in the large area borderd by Solund in the south, Kristiansund in the north and the Jotun Nappe in the east. This review formed the basis for a discussion on models on the ascent of the ultrahigh-pressure (UHP) Western Gneiss Region, and also the role of a mega-shear zone like the Nordfjord–Sogn Detachment Zone. The author drew attention to the following points: **i)** As shown by Young et al. (2007), the Nordfjord area preserves an uninterrupted, gradational transition from high-pressure amphibolite (**~1.5 GPa**) through quartz eclogite (**~2.4 GPa**) to UHP coesite eclogite (**~3.3. GPa**). This shows that the Western Gneiss Region (WGR) was subducted and exhumed as one large body. **ii)** The final amphibolite facies overprint of **0.5 GPa** and **700°C** recorded across the WGR mandates a thermal setting identical to the modern Basin and Range province or the Tibetan

Plateau. These two areas are sites of rapid, large-scale upper crustal extension, indicating that the WGR was effected by the same processes. **iii)** $^{40}Ar/^{39}Ar$ muscovite ages show a consistent decrease from **399 Ma** at the eastern edge of the WGR to **390-385 Ma** in the HP areas at the western edge of the WGR. This east–west gradient was interpreted to be most compatible with exhumation of the UHP rocks as part of a relatively coherent slab, where the eastern areas were exhumed earlier than the western areas that rose from a larger depth. **iv)** These data, combined with other structural and petrological observations, lead Hacker (2007) to a tectonic model in which a coherent, **30 km** thick UHP body was exhumed across the Moho progressively from east to west – meaning that the present eastern parts of the WGR reached shallow crustal levels earlier than the western parts. It was emphasised that this conclusion applies to the WGR treated as a whole, although more complicated imbrication of the WGR had been suggested locally by other workers (e.g. Terry & Robinson 2003).

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An extended summary will be given of Johnston et al. (2007b):

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(52) Johnston et al. (2007b) performed a study that particularly focused on the Nordfjord–Sogn Detachment Zone (NSDZ) in the area between Eikefjord and Hyen (Gloppen), an area which they named the “Hornelen segment” of the NSDZ. The main purpose of the study was to reveal how the development of this zone put constraints on the exhumation of the UHP rocks of the Western Gneiss Region (WGR) and the WGR itself. The NSDZ was reported to have been developed mainly within rocks of the Eikefjord and Lykkjebø Groups (assigned to the Middle Allochthon), but also affecting the minor Svartekari Group (assigned to the Lower Allochthon) at Hyen, as well as the structurally subjacent WGR. The study area also included areas south of Eikefjord.

The study area was bordered as follows: the southern border was defined by the Standal Fault as well as the adjacent WGR south of the fault. The northern border was defined by the Haukå Fault that follows along the southern margin of the Hornelen Devonian Massif. In the west the border line was drawn across the Florø penninsula and the Stavøy island, and to the east the border was going southward from the Sandane region.

Within these borders we also find the Håsteinen Devonian Massif and surrounding area, i.e. the study area of the *present doctor thesis*. However, in their work, Johnston et al. (2007) did not report any data from the study area of the present thesis. Nevertheless, the Eikefjord and Lykkjebø Groups of the NSDZ continue into the study area of the present thesis. Hense, the discussions by Johnston et al. (2007b) on the development of the NSDZ, based on data from the two groups, are relevant to the present thesis – as are also discussions on some other issues. However, the publication of Johnston et al. (2007b) unfortunately arrived in the final stages of the work on the present thesis. Hense, time frames did not allow extensive discussions of the results of Johnston et al (2007b) to be included in the thesis. Nevertheless, some references to the work of Johnston et al. (2007b) are included throughout the thesis.

Although rocks of the NSDZ dominate the study area of Johnston et al (2007b), the Lower Plate and Upper Plate rocks of the detachment system are also present within their study area. The Lower Plate, defined by the WGR, is present in the easternmost parts, near Sandane, continuing to the southern- and southwesternmost parts of their study area, south of the Standal Fault. The Upper Plate, defined by Caledonian Upper Allochthon outboard terranes and overlying Devonian sediments, is present in the west, bordered to the north by the Sunnar fault (merging with the Eikefjord Fault towards the west) and to the south by the Standal Fault. The Standal fault is itself developed within, and cutting, the belt of sheared rocks belonging to the Standal Segment of the NSDZ.

Johnstone et al (2007b) presented data from four different parts of their study area, of which area no. i) is a large area with numerous localities, whereas area no. ii), iii) and iv) are single localities – as can be detailed as follows: **i)** A large majority of the data have been aquired from numerous localities located northeastward from the Eikefjorden area, within the NSDZ. **ii)** A single locality is present at the Stavøya island within the Upper Plate; **iii)** a single locality is found at Standal at the southwesternmost part of the study area, i.e. at the sheared and faulted contact between the Upper Plate and the NSDZ; and **iv)** a single locality is defined by the Naustdal eclogite within the Lower Plate/WGR.

From their study area, Johnston et al. (2007b) obtained data on structural geology, thermobarometry and geochronology (Sm/Nd). Some of their main results and interpretations are referred in the following extended summary:

Structural geolgy:

Event 1: Caledonian contraction, created isoclinal folds in the Lykkjebø and Eikefjord Groups during top-E movements.

Event 2: Formation of the NSDZ , a **2–6 km** thich shear zone with a maximum shear stress of **τ = 28 Mpa** in structurally high parts, and **τ = 24 Mpa** in structurally low parts of the NSDZ.

Event 3: Formation of low-angle ductile-brittle detachments, notably the *Hornelen, Sunnarvik and Standal detachments*, which are meter to decimetre thick fine grained mylonites overprinted by brittle fault cores with pseudotachylite and fault gouge. Juxtaposes Upper Plate rocks with the NSDZ rocks. The authors also introduced, and described the new, westward and shallow-dipping "Blåfjellet detachment", present within the NSDZ near lake Endestadvatnet.

Event 4: Formation of E-W striking, high-angle normal faults and strike-slip faults, notably the *Eikefjord Fault and the Standal Fault*, cutting the ductile-brittle low-angle detachments. *Formation of E-W trending folds* within the Devonian sediments and subjacent units. The folding was interpreted to have accompanied the faulting of event 4.

Thermobarometry:

Sunnarvik Group [Identical to the Høydalsfjorden Complex of the *present thesis*], part of the Upper Allochthon of the Upper Plate: Generally reported to have greenschist facies mineralogy, but one locality at Stavøya island yielded **9.1 +/- 1.3 kbar, 442 +/- 71°C**, indicating local albite-epidote-amphibolite facies and burial to **30 km**. [The locality is situated **10 km** west of the study area of the present thesis].

<u>*Lykkjebø Group*</u>, part of the NSDZ: Ideoblastic peak pressure mineral assemblages yielded **13.3–17.7 kbar, 537–618 °C**, i.e. upper amphibolite facies prograde peak conditions. Overprinted by retrograde top-W asymmetric shear fabrics that yielded **8.5–11.6 kbar, 519–641 °C**, i.e. lower amphibolite to greenschist facies.

<u>*Eikefjord Group*</u>, part of the NSDZ: Peak mineral assemblage gave **16.9–18.0 kbar, 577–582 °C**, i.e. upper amphibolite facies. Overprinted by retrograde top-W shear that yielded **8.3–8.9 kbar, 524–628 °C**; i.e. lower amphibolite to greenschhist facies.

<u>*Western Gneiss Region, the Naustdal Eclogite*</u>: Peak eclogite assemblage: **24.6 +/- 2.1 kbar, 682 +/-73 °C**, i.e. high-pressure (HP) eclogite facies.

<u>*Summary*</u>: Burrial was estimated to be: for the Sunnarvik Group: **30 km**; for the Eikefjord/Lykkjebø Groups: peak: **45–60 km**; and for the WGR: **85 km**. The Eikefjord/Lykkjebø retrograde top-W was reported to have initiated at a depth of **30–40 km**.

Summary of thermobarometry:

Unit	Peak prograde P/T	Retrograde top-W shear P/T
Sunnarvik Group	9.1 +/- 1.3 kbar, 442 +/- 71°C, 30 km	
Lykkjebø Group	13.3–17.7 kbar, 537–618 °C, 45–60 km	8.5–11.6 kbar, 519–641 °C, 30–40 km
Eikefjord Group	16.9–18.0 kbar, 577–582 °C 45–60 km	8.3–8.9 kbar, 524–628 °C, 30–40 km
WGR, Naustdal Eclogite	24.6 +/- 2.1 kbar,682 +/-73 °C, 85 km	

The results were interpreted as follows: Eikefjord and Lykkjebø Groups both experienced prograde peak upper amphibolite facies at **13–18 kbar, 537–618 °C**, and retrograde top-W shear at **8–12 kbar, 519– 641°C**, related to movements on the NSDZ.

Sm-Nd garnet geochronology:

Dating of garnet:

Locality (Lykkjebø Gp)	Garnet core age	Garnet mantle age
Near Svarthumle, east of Eikefjord (J2804L3)	425.1 +/- 1.6 Ma	415.0 +/- 2.3 Ma
Standal (J2805D1)	422.3 +/- 1.6 Ma	407.6 +/- 1.3 Ma
Austre Hydalen, east of Hyen (J2801N)		414.1 +/- 1.6 Ma.

The ages were interpreted in the following way (quote): “Because two-point isochrones cannot test for original homogeneity in $^{143}Nd/^{144}Nd$ among phases or ensure that all phases remained closed to Sm/Nd diffusion, these ages must be interpreted cautiously”. “A maximum age of **425–422 Ma** and a rim age of **415–407 Ma** represent robust ages for the onset and the end of garnet growth during peak prograde amphibolite facies metamorphism in the Lykkjebø Group”.

Strain rates:

The grain size piezometres that were applied to determine stress variations within the NSDZ, (yielding **τ = 28 Mpa** for structurally high levels and **τ = 24 Mpa** for low levels) also yielded data on *differential stress*, that were used to calculate strain rates in the zone. The strain rates were calculated to **$2 \times 10^{-10}s^{-1}$** for structurally high Lykkjebø quartzites, and **$1 \times 10^{-10}s^{-1}$ – $4 \times 10^{-10}s^{-1}$** for the structurally low quartzites. Although comparable to results from other detachment zones, Jonhston et al. (2007b) noted that the strain rates would give unrealistic large shear displacements over the many million years of movements on the NSDZ. However, the authors argued that regardless of the absolute values, the consistancy of quartz grain sizes throughout the shear zone suggests that strain rates were relatively constant, and that strain was initially rather evenly distributed at all structural levels during high-temperature top-W shear along the NSDZ.

Depth and exhumation of the NSDZ

Peak amphibolite facies meatamorphism in the Eikefjord and Lykkejbø Groups was calculated to have developed in the interval **13–18 kbar**. Initiation of top-W shear in the Eikefjord and Lykkejbø Groups, related to movements on the NSDZ, was calculated to have occurred at **8–12 kbar**. Peak conditions in the WGR/Naustdal eclogite was estimated to **25 kbar**. The authors noted that this implied a metamorphic break of **7–12 kbar** from the WGR eclogite to the peak amphibolite facies of the Eikefjord and Lykkjebø Groups.

Timing of allochthon burial, and of development of the NSDZ

The Eikefjord and Lykkjebø Groups constitute the allochthons within which the NSDZ developed. Johnston et al. (2007b) noted that garnet <u>core</u> ages of **425–422 Ma** in the Lykkjebø Group coinside with the post-Wenlockian [Wenlock: **428–425 Ma**] Scandian emplacement of the Solund–Stavfjorden Ophiolite Complex, suggesting that burial of the Middle Allochthon (Eikefjord and Lykkjebø Groups) to depths of **13–18 kbar**/**45–60 km** initiated during ophiolite emplacement. The garnet <u>rim</u> ages of **415–407 Ma** were interpreted to show that the prograde, upper amphibolite facies conditions existed until this time, meaning that these peak conditions spanned the time interval **425–407 Ma**. The authors reminded that the garnet rim ages overlap with the upper range of the **412–400 Ma** ages for formation of the (U)HP eclogites north of Nordfjord. The authors also pointed out that in the Kvamshesten area, $^{40}Ar/^{39}Ar$ muscovite dating of Høyvik Group rocks, analogues to the Lykkjebø Group, has yielded ages around **450 Ma**, (closure temperature **350 +/- 50 °C**) indicating that the Høyvik rocks has *not* been deeply buried during the Scandian phase. The age interval of **425–407 Ma** for the high amphibolite facies in the Eikefjord and Lykkjebø Groups were thus found to correspond to the period of subduction of the WGR to UHP conditions. The datings of Johnston et al (2007)were taken to imply that time brackets could be placed on the movements on the NSDZ: because the garnet ages were seen as representing the peak amphibolite facies assemblages, the subsequent NSDZ shear was interpreted to have developed *later* than **415–407 Ma**, making this the *maximum age* for the shear. $^{40}Ar/^{39}Ar$ muscovite datings in the literature were used to find the *minimum age* for the movements along the NSDZ. The authors claimed that the bulk of the shear occurred during amphibolite facies conditions, and suggested that these movements were essentially completed by the time the rocks had been exhumed to the shallow crustal level corresponding to the muscovite$^{39}Ar/^{40}Ar$-system closure temperature of **350 +/- 50 °C.** The previously published $^{40}Ar/^{39}Ar$ muscovite ages in the area were by Johbnston et al. (2007b) noted to span the interval **402–396 Ma**, and accordingly this was taken as the minimum age for the shear. Using *mean* values for the maximum and minimum estimates, the authors suggested that the top-W ductile displacement along the NSDZ in the study area occurred within the interval **410–400 Ma**. This was taken to indicate that displacement and exhumation associated with the zone were initiated either during, or immediately after, (U)HP metamorphism in the WGR.

Significanse of the NSZD, and a model for the UHP exhumation

These quantitative results were by the authors interpreted to indicate that top-W displacement within the NSDZ exhumed the (U)HP rocks/WGR from the base of the crust, but not from mantle depths. This situation was found to require a three stage model for the exhumation of the WGR: **i)** In the first stage, the (U)HP eclogitic rocks were exhumed from mantle depths of **25 kbar/85 km** to the base of the crust, rising buoyantly along the subduction zone. This exhumation process was believed to have started at **410 Ma**. The authors argued that prior to this, the Eikefjord and Lykkjebø Groups had since **425 Ma** recided at the base of the crust at a depth of **13–18 kbar/45–60 km**, i.e. at upper amphibolite facies conditions. In the suggested model, the Eikefjord and Lykkjebø rocks became underplated by the exhumed (U)HP/WGR rocks, and, hense, the Eikefjord/Lykkjebø and (U)HP/WGR rocks were believed to have been juxtaposed prior to formation of the NSDZ. **ii)** In the second stage, the exhumation of the (U)HP rocks occurred along the NSDZ. The NSDZ itself was reported to have been initiated at **410 Ma**, at a depth of **8–12 kbar/30–40 km**, and the shear zone became penetratively developed within the Eikefjord and Lykkjebø rocks indipendent of tectonostratigraphic contacts. In the model, movements along the NSDZ occurred in the time interval **410–400 Ma**. The authors argued that most of the movements happened during amphibolite facies conditions. The movements were assumed to have ended at **400 Ma**, which is the mean ^{40}Ar/^{39}Ar muscovite age of the sheared rocks, reflecting the gradual exhumation and related cooling affecting the NSDZ itself. **iii)** In the third phase, the WGR/(U)HP eclogitic rocks were exhumed along descrete, decimetre to metre thick, ductile-brittle shear zones – notably the *Hornelen, Sunnarvik and Standal detachments* **–** assumed to have formed after **400 Ma**. The authors suggested that these detachments controlled the formation of the Devonian basins.

Summary

Johnstone et al (2007b) found that top-W shear within the NSDZ did primarily develope within amphibolite facies allochthonous rocks. Thermobarometry and Sm/Nd geochronology was interpreted to show that the top-W fabrics of the NSDZ were initiated at lower crustal depths of **30–40 km** between **410–400 Ma**. The authors argued that the NSDZ was responsible for the postorogenic exhumation of the Norwegian UHP provinces, but only from the base of the crust, not from mantle depths.

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Chapter 2

CHAPTER 2 REGIONAL SETTING, WITH COMMENTS ON SOME CURRENT GEOLOGICAL MODELS

The Caledonian Orogeny reached its culmination in Scandinavia during the Silurian–Devonian collision between the Baltic and Laurentian cratons (Scandian orogeny), marking the final closure of the Iapitus ocean. The orogeny led to full scale imbrication of the continental crust with subduction of the Baltic margin below Laurentia, and emplacement of thrust nappes of outboard, marginal and continental terranes onto the peneplained Baltic platform. These thrust nappes have been ranked into Lower, Middle, Upper and Uppermost Allochthons (Roberts & Gee 1985) that overlie the autochthonous/parautochthonous cratonic gneisses, e.g. the Western Gneiss Region (subducted Baltic margin). In Early Devonian times, the collision phase ceased, and the following collapse of the orogenic welt was accompanied by crustal-scale extension. This led to deposition of Devonian Old Red Sandstone sediments on a western Upper Plate, and isostatic uplift of an eastern and partly subjacent Lower Plate — the Western Gneiss Region. (Andersen et al. 1991; Fossen 1992, 1998, 2000; Milnes et al 1997; Andersen 1998).

The present thesis provides data from both the contractional and extensional history of the Caledonian orogenic belt as apparent in the study area, but with main focus on the extensional development.

2.1 OVERVIEW

Chapter 2 deals with the main geological relationships in the areas of western Norway surrounding the Håsteinen study area. Emphasis will be on the rock architecture found in the coastal areas between Sognefjorden (southern boundary), Nordfjord (northern boundary) and the "Faltungsgraben" (eastern boundary) (see **Fig. 2.1**, **2.2** and **2.3**). A review of the Caledonides of western Norway between Stavanger and Kristiansund has been given by Bryhni & Sturt (1985). The ranking of the tectonostratigraphical units into Autochthons — and Lower, Middle and Upper Allochthons — and their areal distribution, are shown in **Fig. 2.1**. Each of these tectonostratigrahic units may contain several smaller nappes and/or rock complexes. The names and locations of such individual nappes and rock complexes between Stavanger and Nordfjord are shown in **Fig. 2.2**. On this figure, note the usage of the following geographical names:

Nordfjord: name of a fjord, and also of the region along it (outer/inner Nordfjord).

Sunnfjord: name of a region only (outer/inner Sunnfjord).

Sognefjorden: name of a fjord only.

Sogn: name of the region along Sognefjorden (outer/inner Sogn).

Note also the geological names used for the gneiss area located west of the Faltungsgraben:

Western Gneiss Region: Used throughout the text; area containing rocks of the

Western Gneiss Complex: Used on Figures of Chapter 2, and on Plate 1.

Regarding the Devonian deposits, the following terms are used in the present chapter (and throughout the thesis):

Devonian basin: Term used to denote the deposits as they were at the time of deposition or soon after, when the basin still had its original size and defined a depression in the landscape.

Devonian massif: Term used to denote the deposits as they appear at the present day, where they are merely erosional remnants of the original basins, and tend to define mountains, i.e. massifs, in the landscape.

When summarising, in the thesis, literature that use the term “basin” for both the original and present-day deposits, the term may be used in the similar way in the thesis.

Several geological bedrock maps of regional and national scale cover the discussed area: good overviews of the area between Sognefjorden and Nordfjord are offered by the **1:250.000** map of Måløy (Kildal 1970). The relationship of this area, to the general geology of Norway appears on the **1:1 million** "Bedrock map of Norway" (Sigmond et al. 1984); the relationship to offshore areas is found on the **1:3 million** "Bedrock map of Norway and adjacent ocean areas" (Sigmond 1992); and the tectonostratigraphic units appear on the "Scandinavian Caledonides — Tectonostratigraphic map" (Gee et al. 1985).

The coastal areas situated between Sognefjorden, Nordfjord and the "Faltungsgraben" generally contain three major rock units (See **Fig. 2.3**: The unit-numbers given below are shown in the figure):

1. the Middle (?) Devonian continental Old Red Sandstone massifs of Hornelen, Håsteinen, Kvamshesten and Solund, unconformably overlying the Caledonian nappes.
2. the Caledonian nappe pile, consisting of ophiolites and associated melanges, and Precambrian meta-igneous rocks (gneisses) with supracrustal cover sequences, etc.
3. the Precambrian Baltic craton, consisting of basal migmatitic gneisses, granites and meta-supracrustals.

In the following discussion, two geographical areas will be in focus (for location, see **Fig. 2.4** and **2.5**):

Area 1: the area situated between Førdefjorden (south boundary), and a line (north boundary) following Norddalsfjorden and further ENE- and NE-wards along the southern margin of the Hornelen Devonian Massif, to Hyen and inner Nordfjord. Area 1 encompasses the study area.

Area 2: the area situated between Dalsfjorden (south) and Førdefjorden (north). Area 2 is particularly well known due to extensive recent research, and contains rock units that have been correlated with units in Area 1. Before discussing individual units, the detailed tectonostratigraphy for these two areas will be summarised:

Area 1: The area between Førdefjorden and Norddalsfjorden/inner Nordfjord (Sect. 2.3) contains the following 5 major geological units (here listed with youngest division on top) (See **Fig. 2.4**, lowermost box with tectonostratigraphic column; and **Table 2.1**; both containing the unit-numbers listed below; and **Fig. 2.5**):

1. the Håsteinen Devonian Massif (HDM): ~Middle Devonian Old Red Sandstone (ORS) deposits, subjected to syn- or post-sedimentary deformation and metamorphism, resting unconformably on:
2. the Høydalsfjorden Complex (HC): Lower Palaeozoic rocks, previously correlated with the cover sequence to the Solund–Stavfjorden Ophiolite Complex (dated to **443 +/- 3 Ma**), part of the Caledonian nappe pile, emplaced (obducted) during the Scandian continent-continent collision, assigned to the Upper Allochthon, and lying with a steep fault contact towards (but tectonostratigraphically above):
3. the Lykkjebø Group: meta-sediments, probably of continental margin type and possibly an Eocambrian "sparagmite" equivalent. Considered to be part of the Fjordane Complex. Suggested by Andersen & Jamtveit (1990) to be part of a “Middle Plate” that was claimed to exist between the Upper and Lower Plate of the Nordfjord–Sogn Detachment Zone (NSDZ). Assigned to the Lower Allochthon by Bryhni & Sturt (1985), but assigned to the Middle Allochthon by Andersen & Jamtveit (1990) as a cover sequence to:
4. the Eikefjord Group: "Jotun-kindred" meta-igneous rocks of Precambrian origin (maximum **~1700 Ma**). Also suggested by Andersen & Jamtveit (1990) to be part of the “Middle Plate” that they claimed should exist between the Upper and Lower Plate. Grouped as part of the Fjordane Complex, assigned to the Middle Allochthon by Bryhni & Sturt (1985) as well as Andersen & Jamtveit (1990), correlated with the Jotun Nappe Complex, positioned tectonostratigraphically on top of:
5. the Jostedalen Complex: migmatitic and granitic basal gneisses, part of the Western Gneiss Region, assigned to the Autochthon/Parautochthon, forming the lowest unit in the tectonostratigraphy of western Norway.

Area 2: The area between Dalsfjorden and Førdefjorden (Sect. 2.5) contains the following tectonostratigraphy (again listed with youngest division on top). (See **Fig. 2.4**, uppermost box with tectonostratigraphic column; and **Table 2.2**; which both contain the unit-numbers below):

1. the Kvamshesten Devonian Massif, lying unconformably on:
2. the Stavenes Group (primary/oceanic sedimentary cover to the Solund–Stavfjorden Ophiolite Complex), lying unconformably on:
3. the Solund–Stavfjorden Ophiolite Complex (S-SOC), previously dated to **443 +/- 3 Ma**., lying with thrust contact (and later extensional fault contact) against:
4. the Sunnfjord Melange (obduction-melange below the S-SOC), lying with thrust, and partly unconformable, contacts towards:
5. the Herland Group (continental margin sediments, laterally equivalent to the Stavenes Group), lying unconformably on:
6. the Høyvik Group (continental margin sediments, probably of Eocambrian age), lying unconformably on:

7. the Dalsfjord Suite (Precambrian igneous rocks of the "Jotun-kindred", previously dated to **1634 +/- 3 Ma**), lying with thrust and later extensional detachment/fault contact towards:
8. the Askvoll Group (extensively mylonitized meta-sediments, meta-volcanics and meta-intrusives; first assumed to be of Caledonian age, now interpreted to be of Early/Mid Proterozoic age based on previous dating of a quartz diorite to **1640.5 +/- 2.3 Ma**, interpreted to be part of the Western Gneiss Complex), lying with mylonitized contact towards:
9. the Western Gneiss Region proper (migmatitic and granitic gneisses forming the lowest unit in the tectonostratigraphy of western Norway).

Since the interpretation of the Devonian massifs – including Håsteinen – depends on the correlation and interpretation of these sub-Devonian tectonic and lithological units in a region of varied and complicated geology, the following sections of this chapter include short descriptions of each unit, indicating the relevant literature. The first section discusses the Western Gneiss Region, which is common to both areas (Sect. 2.2) and which has been subjected to confusing changes in nomenclature. Then follows a summary, in two parts, of the tectonics of **Area 1** between Førdefjorden and Norddalsfjorden/inner Nordfjord, i.e the surroundings of the Håsteinen Devonian Massif (Sect. 2.3 and 2.4); succeeded by a presentation of the tectonics of the important area to the south, **Area 2**: Dalsfjorden–Førdefjorden (Sect. 2.5). The next sections present a synthesis of the Lower Palaeozoic units (Sect. 2.6) and the Devonian massifs (Sect. 2.7) for the whole region, before finally discussing the *models* proposed for the late- or post-Caledonian tectonic processes of the region (including the formation and deformation of the Devonian massifs): the "Solundian Orogeny" model (Sect. 2.8), and the "detachment" model (Sect. 2.9).

2.2 WESTERN GNEISS REGION

2.2.1 GENERAL GEOLOGY

The Western Gneiss Region (WGR) is located between the "Faltungsgraben" (Central Trough) (**Fig. 2.1** and **2.2**) in the east, and the coast in the west – and between Bergen in the south and the Kristiansund area in the north (Bryhni & Sturt 1985; Bryhni 1989; Skår 1998). The reader is reminded that the names "*Western Gneiss Region"* and "*Western Gneiss Complex"* are referring to the same area.

The rocks in the region are very characteristic and can be recognised over large areas. Typically the rocks are granitic to granodioritic gneisses and migmatites with red feldspar augen and frequent veins of pegmatite and aplite. In addition, dolerite (gabbro) and amphibolite are present in places (Bryhni 1977; Bryhni et al 1981; Bryhni 1989; Skår 1998). More special rock types occur locally all over the gneiss region. These are granulites, anorthosites, eclogites and ultramaphites. The anorthosites have been given varying interpretations: Bryhni (1977) interpreted anorthosites as intrusive in the gneisses. At Måløy, Kildal (1970) interpreted such rocks as being thrust nappes. However, this interpretation was not confirmed by Krabbendam & Wain (1997), who viewed the rocks as an integral part of the gneisses. The eclogites between Sogn and Trøndelag are found to be associated with all these rocks (e.g Griffin & Mørk 1981; Terry et al. 2000a, 2000b; Engvik & Andersen 2000). The eclogites have been interpreted as formed ***in situ*** (Griffin et al. 1985), implying that they are not of exotic origin, but part of the Baltic craton. The region also contains various supracrustal rocks assumed to be of Eocambrian or Precambrian ages (e.g. Bryhni 1989).

The ages of the rocks cover a wide time span. Beyer et al. (2004) report that Archaean crustal rocks have not been found in the WGR, but that bodies of peridotites in the WGR had an Archaean pre-crust origin, and that these rocks were brought into the WGR in Proterozoic times. U/Pb ages from zircons and Rb/Sr whole rock ages from the crustal rocks of the Western Gneiss Region yield dates ranging essentially between **1800–1500 Ma**. (Brueckner 1972; Griffin & Mørk 1981; Griffin and Brueckner 1985; Gebauer et al. 1985; Bryhni & Sturt 1985; Kullerud et al.1986; Skår 1998; Austrheim 2003; Skår & Pedersen 2003), and are interpreted as the age of the protoliths. The area also contains granites yielding ages around **900 Ma** (e.g. Milnes et al. 1988; Skår 1998; Skår & Pedersen 2003), related to the Sveconorwegian reworking. Early Palaeozoic, Caledonian reworking caused by the Laurentian–Baltic collision, is evident from the fact that Sm/Nd and U/Pb datings of eclogites (Griffin & Brueckner 1980, 1985; Kullerud et al. 1986) give essentially Early Palaeozoic ages, as do $^{40}Ar/^{39}Ar$, K/Ar and Rb/Sr ages (e.g. Bryhni et al. 1971; Lux 1985; Kullerud et al. 1986; Cuthbert 1991; Chauvet & Dallmeyer 1992; Fossen & Dunlap 1998; Andersen 1998). The Early Palaeozoic/Caledonian reworking was reported already by Holtedahl (1936), who observed that deformation of Early Palaeozoic rocks had also affected the subjacent gneisses. In recent years, considerable attention has been given to the study of ultra-high-pressure (UHP) and high-pressure (HP) rocks within the WGR, in an attempt to refine the timing and rate of the subduction and exhumation of this western part of Baltica during the Scandian phase of the Caledonian orogeny. This is further discussed in Sect. 2.8.

The gneiss region is assumed to represent the lowest tectonostratigraphic level exposed in western Norway. During the research history, the tectonic status of the region has been disputed, and at least four different models have been presented. The gneisses have been regarded as : **a)** a micro-continent welded to the baltic shield by a Caledonian “Jotunheimen suture” (Banham et al. 1979); **b)** a gigantic nappe possibly even overlying low grade sediments (Mykkeltveit et al.1980; Hossack in Bryhni et al. 1981), **c)** a rock complex that have a parautochthonous position (Gee 1975, 1980), and finally, **d)** an area that has a complete continuity with the Fennoscandian craton (Corfu 1980; Milnes et al. 1997; Skår1998; Skår & Pedersen 2003). The last model is now supported by most workers, and the general recognition of the eclogites being of late Caledonian age has led to general acceptance of a model where the Western Gneiss Region – representing a continious part of the Baltic craton – were subducted below the Laurentian craton (Cuthbert et al. 1983; Grifin et al. 1985; Andersen et al. 1991, Milnes et al. 1997; Hacker 2007).

To the east, the WGR **(Fig 2.1** and **2.2)** is overlain by the major Caledonian Jotun Nappe Complex (and south of this complex: the Bergsdalen Nappes) with its subjacent sheared meta-sediments of the décollement zone (Milnes & Koestler 1985; Fossen 1992; Milnes et al. 1997). In the western areas, i.e. along the coast, the WGR is overlain by the Caledonian nappe pile, but separated from these nappes by the so-called Nordfjord–Sogn Detachment Zone (Norton 1986) (Sect.2.8), and its southward continuation, the Bergen Arc Shear Zone (BASZ) (Wennberg et al. 1998).

2.2.2 NOMENCLATURAL PROBLEMS

During the last decades, no consensus has been reached as to how the rocks between the coast and the "Faltungsgraben" should be grouped and named, and this problem still prevails. The WGR occupies a large area of very complex and poorly known geology, and much work remains before it will be satisfactorily understood. In the following, the nomenclature problems arising from this fragmentary knowledge of the WGR will be illustrated from the area located between Sognefjorden, Nordfjord and the "Faltungsgraben" (**Fig. 2.2**, **2.3** and **2.4**), with particular focus on the area between Førdefjorden and inner Nordfjord (**Fig. 2.5**).

Since the 1960s, the publications authored by Bryhni has been of particular importance in the naming of these gneiss areas. Bryhni (1966) presented the first modern attempt to group the rocks between the coast and the "Faltungsgraben". The basal migmatitic and granitic gneisses in the *east* were assigned to the **Jostedalen Complex** (Bryhni 1966; Bryhni & Grimstad 1970; Bryhni 1977; Bryhni 1989). The rocks situated mainly to the *west*, such as the metasediments of assumed Eocambrian age; the "metamorphic supracrustals" (within the Jostedalen gneisses); the "anorthosite-kindred" rocks; and the Lower Palaeozoic rocks, were all assigned to the **Fjordane Complex**. (The

"Faltungsgraben" itself contains, from bottom to top, Cambro–Silurian rocks in the basal décollement zone, Valdres Sparagmites of assumed Eocambrian age, and the igneous rocks of the Jotun Nappe. The latter two constitute the Jotun Nappe Complex, with the Sparagmites as an inverted cover). The idea behind the subdivision into the Jostedalen and Fjordane complexes was that the rocks of the Fjordane Complex, situated west of the Jostedalen Complex, were considered eqvivalent to the rocks in the "Faltungsgraben", situated east of the Jostedalen Complex – but that the western Fjordane Complex rocks were much more deformed than the eastern “Faltungsgraben” rocks. The Jostedalen Complex was supposed to underlie the two other units. It should be noted that in this original model, ***all*** supracrustals and Caledonian nappes along the coast, except the Devonian "Old Red Sandstones", were assigned to the Fjordane Complex. Later, this first model has been modified due to the increased complexity arising from the accumulation of new information during the last decades, and the areal extent of the two complexes has been changed significantly (see below).

Both the Jostedalen and Fjordane Complexes contained gneisses (Bryhni 1966), and after this first grouping attempt, different authors used various names for the area covered by the two complexes. Bryhni (1966) did not use all-embracing terms like the "Western Gneiss Region/Complex" in connection with his original Fjordane/Jostedalen subdivision. Bryhni et al. (1971), however, used the term "Gneiss Region", meaning all rocks between the "Faltungsgraben" and the west coast except the ORS. Bryhni (1977) still used the term "Gneiss Region" for the areas west and northwest of Jotunheimen, but for the first time the Lower Palaeozoic rocks along the coast were apparantly not included in the "Gneiss Region". Brueckner (1979) used the term "Basal Gneiss Region" to indicate the same area as Bryhni (1977). Eventually, Bryhni et al. (1981) introduced the term "Western Gneiss Region" (WGR), once again meaning all rocks in the area between the "Faltungsgraben" and the west coast except the Devonian deposits. The name “Western Gneiss Region” has subsequently been extensively used in the literature, although variations have occurred as to what rocks are actually included in the "Western Gneiss Region", by the different authors.

Over the years, the extent of the Jostedalen Complex has been considerably expanded westwards at the expence of the Fjordane Complex, compare e.g. Bryhni (1966) and geological 1:50 000 maps of Bryhni & Lutro (1991a, 1991b, 2000a, 2000b). (*Note that these maps have been revised by Bryhni & Lutro 2000a, 2000b, where the gneisses – after yet another name change – are assigned to a group called “Basement, mainly Precambrian rocks”, instead of the previous “Jostedal Complex*). The westward expansion of the Jostedalen Complex means, for example, that the rocks to the south of the Håsteinen area, i.e. between the Standal Fault and Førdefjorden, have been changed from being part of the Fjordane Complex (Bryhni 1966) to be part of the Jostedalen Complex (see maps of Bryhni & Lutro 1991a, 1991b, 2000a, 2000b) (**Fig. 2.3**). The Jostedalen Complex will not be further discussed here, but the Fjordane complex is further treated below.

The Fjordane Complex between Førdefjorden and Inner Nordfjord (the Eikefjord–Gloppen area) (**Fig. 2.4** and **2.5**) contains metamorphic supracrustals and "anorthosite-kindred" rocks. Rocks of this type are located at several places in the areas between the "Faltungsgraben" and the west coast. The following areas can be mentioned as examples:

a) the Oppdal area

b) the Lesjaskogvatn area

c) the Tafjord–Grotli–Sotaseter area

d) the Eikefjord–Gloppen area

The rocks at a) are situated close to the Trondheim Nappe Complex, and the rocks at b) and c) (**Fig. 2.1**) are located within the area comprising basal migmatitic gneisses of the Jostedalen Complex. Focus will here be on the Fjordane Complex of the d) Eikefjord–Gloppen area **(Fig. 2.5)**, which is also represented within the study area of the present thesis.

The Fjordane Complex was interpreted by Bryhni et al. (1981) and Bryhni (1989) to lie tectonostratigraphically on top of the Jostedalen Complex. The igneous rocks of the Eikefjord–Gloppen area may be lithologically correlated with the Dalsfjord Suite and the Jotun Nappe Complex.

Bryhni et al. (1981) introduced a further subdivision of the Fjordane Complex into what they called the "Lower Part" **(Table 2.1)** (see also Bryhni et al. 1971). A corresponding "Upper Part", however, was not mentioned, although it is possible that the Early Palaeozoic rocks along the coast were thought to constitute an "Upper Part" **(Table 2.1**). This assumption is based on the fact that, in Bryhni et al. (1981), the Early Palaeozoic rocks were for the first time treated separately from the WGR, although they were still considered to be part of the WGR. (But, the separate treatment may be seen as the first step in the process of “removing” the Early Palaeozoic rocks from the Fjordane Complex of the Western Gneiss Region).

In the area between Inner Nordfjord and Eikefjorden, Bryhni et. al. (1981) furthermore divided the "Lower Part" into 2 units; the Lykkjebø Group, consisting of supracrustals (meta-psammites and meta-pelites with carbonate), and the Eikefjord Group consisting of ortho-gneisses of the "anorthosite-kindred" (**Fig. 2.4** and **2.5**). The term "Lower Part" was later abandoned by Bryhni (1989), whereas the terms "Eikefjord" and "Lykkjebø Groups" have been retained.

Bryhni et al. (1981) and Bryhni (1989) noted that meta-igneous rocks of the Eikefjord Group can be found as layers within the meta-sedimentary rocks of the Lykkjebø Group, and vice versa, and left open the question as to whether one of the groups is tectonostratigraphically on top of the other. The contacts between the "metamorphic supracrustals" (Lykkjebø Group) and the "anorthosites" (Eikefjord Group) of the Eikefjord–Gloppen region are not clearly defined (Bryhni et. al. 1981; Bryhni 1989). This is due to the presence of interfolding and high shear-deformation in the area, which have turned the rocks into mylonites and ultramylonites and caused internal mixing. In the Eikefjord area, the southern boundary of the Eikefjord/Lykkjebø rocks, i.e. towards the Early Palaeozoic rocks (**Fig. 2.5**), is defind by the steep, brittle *Sunnar Fault*, which is trending east-southastwards from Sunnarviken (Bryhni et al. 1981; and this work). (*Note that the name “Sunnar Fault” has been changed to “Sunnarvik Fault” on the 1:50 000 maps of Bryhni & Lutro 2000a, 2000b, a name also adopted by Johnston et a. 2007b. However, the original name “Sunnar Fault” will be used throughout the present thesis*).

Bryhni and Sturt (1985) "reorganised" the tectonostratigraphic “position” of the rock units of the Fjordane and Jostedalen Complexes of the Western Gneiss Region (i.e. including the Eikefjord and Lykkjebø

Groups) in order to conform to the system of regional allochthons used in the Scandinavian Caledonides **(Fig. 2.1, Table 2.1)**. The rocks of the Jostedalen Complex in the eastern part of the Western Gneiss Region were assigned to be in Autochthonous/Parautoch-thonous position. In the western areas, between the fjords of Nordfjord and Sognefjorden, the areal extent of the "Western Gneiss Region" was, however, significantly reduced compared to the earlier publications by Bryhni (e.g. Bryhni 1966; Bryhni et al. 1981; Bryhni 1989). This area-reduction was obtained by excluding the entire former Fjordane Complex from the "Western Gneiss Region" (Bryhni & Sturt 1985). Instead, the former Fjordane Complex-rocks of Bryhni (1966) and Bryhni et al. (1981) between Førdefjorden and Sognefjorden were assigned to the Autochthon/Parautochthon and denoted "metamorphic supracrustals of uncertain position" (Bryhni & Sturt 1985) (**Fig. 2.1**). The former Fjordane Complex-rocks of the Eikefjord–Gloppen area were also reorganised, but differently: The meta-sedimentary Lykkjebø Group was assigned to the Lower Allochthon as a cover sequence, and the meta-igneous Eikefjord Group was placed on top as part of the Middle Allochthon and correlated with the Jotun Nappe and the Dalsfjord Nappe (Bryhni & Sturt 1985) (**Fig. 2.1**). The Early Palaeozoic rocks along the coast (including the Høydalsfjorden Complex, as defined in the present thesis) were assigned to the Upper Allochthon. If we now compare with the Dalsfjord Nappe to the south of Førdefjorden, a naming paradox appears to be present: the cover of meta-psammites and other sediments on top of the Dalsfjord Suite was interpreted as part of the Middle Allochthon, whereas similar supracrustals in the Lykkjebø Group in the Eikefjord–Gloppen region (**Fig. 2.4** and **2.5**) were – as we have seen above – for unexplained reasons instead assigned to the Lower Allochthon.

It should furthermore be noted that the "metamorphic supracrustals" and the "anorthosite-kindred" rocks of the Tafjord–Grotli–Sotaseter area (**Fig. 2.1**), which according to Bryhni (1977) are analogous to the Lykkjebø and Eikefjord Groups with respect to both lithology and tectonostratigraphic position, have ***not*** in a similar way been assigned to Lower and Middle Allochthon positions, but are instead considered to have an "uncertain position", although assigned to the Autochthon/Parautochthon and geographically located within the Western Gneiss Region (Bryhni & Sturt 1985).

In a later publication (Bryhni 1989), the "Fjordane Complex" is once again included in the WGR, and defined as containing the Lower Palaeozoic rocks in the west, the metamorphic supracrustals (= Lykkjebø Group) and the "anorthosite Kindred"-rocks (= Eikefjord Group), i.e. the same usage as in Bryhni et al. (1981).

These naming problems illustrate the difficulties encountered when the "allochthon-system" is applied to the very complex and poorly known "Western Gneiss Region".

The papers of Andersen et al. (1990) and Furnes et al. (1990) contained figures with regional maps suggesting that the meta-anorthosites, etc. of the Eikefjord Group may be correlated with the igneous rocks of the Dalsfjord Suite which constitutes most of the Dalsfjord Nappe. Accordingly, these authors correlated the meta-sediments of the Lykkjebø Group (**Fig. 2.4** and **2.5**) with the meta-sediments of the Høyvik Group which lies unconformably on the Dalsfjord Suite (**Fig. 2.4**). Swensson & Andersen (1991) make the same correlations on *their* regional map figure, and they assign this unit of continental "basement and cover" to the Middle Allochthon position, thus opposing Bryhni & Sturt (1985) who assigned the meta-sediments to the Lower Allochthon. Most later publications from these areas have correlated the Lykkjebø Group with the Høyvik Group, and the Eikefjord Group with the Dalsfjord Suite (e.g. Osmundsen & Andersen 2001).

2.3 TECTONOSTRATIGRAPHY BETWEEN FØRDEFJORDEN AND NORDDALSFJORDEN / INNER NORDFJORD

The present section deals with the tectonostratigraphy of the area between Førdefjorden and Norddalsfjorden/inner Nordfjord (**Fig. 2.4** and **2.5**) (an area which includes the Håsteinen study area), and starts with the lowest level.

2.3.1 JOSTEDALEN COMPLEX

The general geology of the basal gneisses was described above (Sect. 2.2). In the area considered here, the rocks of the Jostedalen Complex (**Fig. 2.3**) are situated between Førdefjorden and the Standal Fault (see maps of Andersen & Jamtveit 1990; Swensson & Andersen 1991; Bryhni & Lutro 1991a, 1991b, 2000a, 2000b), an area which is located to the south of the Høydalsfjorden Complex and the Håsteinen Devonian Massif (**Fig. 2.4**). On regional maps in more recent publications, this area is not called the Jostedalen Complex, but instead the Autochthon/Parautochthon (e.g. Hartz & Andresen 1997; Skår & Pedersen 2003), or simply the Western Gneiss Region (e.g. Andersen 1998; Fossen & Dunlap 1998; Krabbendam & Dewey 1998; Milnes and Koyi 2000; Schärer & Labrousse 2003; Johnston et al. 2007b). The rocks consist of granitic augen gneisses, amphibolites, quartz-feldspathic (meta-arkosic) gneisses with zones of meta-quartzite and mica-schist, and eclogites (Kildal 1970), together with gneisses of granodioritic to granitic composition and bodies of ultramafics (Bryhni & Lutro 1991a, 1991b, 2000a, 2000b). The area has earlier been assigned to the Fjordane Complex (Bryhni 1966), the Askvoll Group (Skjerlie 1969; Furnes & Skjerlie 1972), the Vevring Complex (Skjerlie & Tysseland 1981), the Jostedalen Complex (Swensson & Andersen 1991), and to the Naustdal Group of the Jostedalen Complex (Bryhni & Lutro 1991a, 1991b, 2000a, 2000b). The northern part of the area, which defines a belt of mylonites, was interpreted by Andersen & Jamtveit (1990) as the "Standalen Detachment", a segment of the Nordfjord–Sogn Detachment Zone (**Fig. 2.3**) (Sect. 2.8). It should be noted that the *Standal Fault* is a brittle fault which follows along the more than **2 km** thick ductile mylonites of the "Standalen Detachment Zone", and which cuts the mylonitic fabric. The structural geology of the westernmost exposure of the mylonite zone and fault, has recently been studied in detail by Krabbendam & Dewey (1998); and Andersen (1998) has reported an $^{40}Ar/^{39}Ar$ muscovite plateau age of **398 Ma** from the same area. Also in the same area, on the south side of the Standal Fault, Johnston et al. (2007b) obtained a Sm/Nd garnet *core* age of **422.3 +/- 1.6 Ma** and a *rim* age of **407.6 +/- 1.3 Ma**. Thermobarometry from the same locality yielded peak prograde metamorphism of around **15 kbar, 539 +/- 75 °C**, and retrograde metamorphism of around **9.6 kbar, 525 +/- 70 °C**. The latter authors also investigated fault striations related to the ductile-brittle Standal Fault

2.3.2 EIKEFJORD GROUP

2.3.2.1 DESCRIPTION

The ortho-gneisses of the Eikefjord Group include meta-anorthosites, syenitic to monzonitic biotite-epidote gneisses, amphibolites, meta-gabbros and ultramafites. A petrographic study of these rocks and the rocks in the Lykkjebø Group (see below) was carried out by Bryhni (1958). The Eikefjord Group is essentially present in the area between Eikefjord and Gloppen **(Fig. 2.5)** (Bryhni et al. 1981). Remnants of a Proteozoic granulite facies mineralogy was reported from some of the rocks (Bryhni 1958, 1989; Bryhni et al. 1981; Johnston et al. 2007b). A thin belt of rocks along the southern side of the Standal Fault has been assigned to the Eikefjord Group by Bryhni & Lutro (1991a, 1991b, 2000a, 2000b) (**Fig. 2.5** and **2.6**). The same rocks were also reported to constitute the peninsula reaching from Eikefjord and westwards to the town Florø (**Fig. 2.4**) (Bryhni et al. 1981), an area which Skjerlie & Tysseland (1981) assigned to their Vevring Complex. Rocks belonging to the Eikefjord Group are present in the northern part of the study area of the present thesis (see Ch. 3).

The meta-igneous rocks (meta-anorthosites, etc) of the Eikefjord Group are generally – by correlation with the Jotun Complex – presumed to have originated around **1600-1700 Ma**. (Bryhni & Sturt 1985). Abdel-Monem & Bryhni (1978) have dated meta-anorthosites of the Eikefjord area, belonging to the Eikefjord Group, by the Rb/Sr method, whereby **7** rock samples from Sande in Gloppen and from Eikefjord have been combined into a common regression line giving a 7 point "isochron" of **1511 +/- 64 Ma** with intercept at **0.7040 +/- 0.0002** . The age was interpreted as a minimum age of intrusion.

The rocks have been structurally and metamorphically reworked several times, latest during penetrative mylonitisation under amfibolite to greenschist facies conditions, but Bryhni (1989) and Johnston et al (2007b) nevertheless reported remnants of a Proterozoic granulite facies mineralogy in lenses of little deformed rocks.

The penetrative mylonite fabric frequently overprints and obliterates the older structures, and Andersen & Jamtveit (1990) interpreted this fabric as part of the extensive mylonites defining the Nordfjord–Sogn Detachment Zone. Wilks & Cutbert (1994) also defined the rocks in the Eikefjord–Gloppen area as a top-to-the-west shear zone, but called it the ”principal shear zone”. Also Johnston et al (2007b) supported the view that the shear is penetratively developed in the Eikefjord and Lykkjebø Groups. Wilks & Cuthbert (1994), who carried out detailed structural analyses in the Eikefjord–Gloppen mylonites, expressed no support to the proposal made by Andesen & Jamtveit (1990) that the Eikefjord–Gloppen area represents a separate “Middle Plate” situated between the well established Upper Plate and Lower Plate. Johnston et al. (2007b) did not either use the concept of a “middle plate”.

The Eikefjord Group rocks were originally grouped as part of the Fjordane Complex, which were believed to be situated on top of the Jostedalen Complex (Bryhni 1989). According to Bryhni et al. (1981) and Bryhni (1989), extensive interfolding and intershearing of the Eikefjord Group rocks with rocks of the Lykkjebø Group have made it extremely difficult to determine the mutual tectonostratigraphic position for these two groups. Bryhni & Sturt (1985) nevertheless assigned the Eikefjord rocks to the Middle Allochthon – as was also done on the map-figure in Andersen & Jamtveit (1990). The rocks in the Eikefjord Group have been correlated with the Dalsfjord Suite, located between Dalsfjorden and Førdefjorden (see map-figure in Andersen & Jamtveit 1990), and with the Jotun Complex. On the map-figure in Furnes et al. (1990), the rocks of the Eikefjord Group are *included* in the Dalsfjord Suite. Johnston et al (2007b) assighned both the Eikefjord and Lykkjebø Groups to the Middle Allochthon.

2.3.2.2 THE EIKEFJORD FAULT

In the Eikefjord Valley area, the ***northern*** boundary of the Eikefjord Group is defined by the steeply north-dipping, E-W trending, brittle <u>Eikefjord Fault</u> that separates the Eikefjord Group to the south from the Lykkjebø Group to the north (**Fig. 2.4** and **2.6**) (Bryhni et al.1981; Johnston et al. 2007b; and this thesis). Towards the east, the fault apparently does not merge with the Standal Fault (**Fig. 2.6**) (Kildal 1970; Johnston et al. 2007b). The Eikefjord Fault is briefly described in Sect. 3.2 (under “Boundaries”).

It is necessary to draw attention to an unfortunate use of the name “Eikefjord Fault” in a recent publication: On the map-figure in Braathen et al. (2004), the name “Eikefjord Fault” is not used in the traditional way, i.e. to denote the fault along the ***northern*** margin of the Eikefjord Group in the Eikefjord Valley, but is instead used for the fault that defines the ***southern*** margin of the Eikefjord Group in the valley, a fault which in numerous publications bears the name the “<u>Sunnar Fault</u>” (or “Sunnarvik Fault” on the 1:50 000 maps of Bryhni & Lutro 2000a, 2000b; and in the paper of Johnston et al. 2007b). The “transfer” of the name “Eikefjord Fault” from the northern to the southern boundary of the Eikefjord Group, may cause unwanted confusion. In the present thesis, the traditional use of the fault names will be maintained.

2.3.2.3 DOLERITE DYKES IN THE EIKEFJORD GROUP

At least **30** individual N-S trending and vertical dolerite dykes with a continental tholeiitic composition have been reported from Sunnfjord (Reusch 1881; Kolderup 1928; Kildal 1970; Skjerlie & Tysseland 1981;Torsvik et al. 1997), and a large number of them are located in the Eikefjord Group rocks in the peninsula between Florø and Eikefjord and nearby areas to the south and west (Bryhni & Lutro 2000a; Lutro & Bryhni 2000).

Skjerlie & Tysseland (1981) interpreted the dykes to have intruded in the Lower Devonian. This was based on the fact that the dykes cut all Caledonian structures but stop abruptly against the Devonian deposits, attesting to a pre-Middle Devonian age. Torsvik et al. (1997) showed that this was not correct, and used palaeomagnetic dating to obtain a Permian, c. **270–250 Ma**, age for the dyke intrusions. This age is similar to K-Ar ages obtained from dykes at Sotra west of Bergen, which yielded **262± 6 Ma** (Løvlie & Mitchell 1982). The oldest dykes from Sunnhordland gave corresponding ages at **280–260 Ma** (K-Ar ages, Færseth et al. 1976), or **250–260 Ma** (^{40}Ar/^{39}Ar ages, Fossen & Dunlap 1999). All these dykes may thus be related to the Permo-Triassic tectonic evolution of the Norwegian continental shelf.

In addition to the above dykes, two peralkaline, ultrapotassic syenite dykes of Middle Permian age, **261±6** / **260±2** and **256±6** / **256±6** (K-Ar ages, Furnes et al. 1982), have been found at Heilefjellet north of Dalsfjorden, within the "Jotun rocks" of the Dalsfjorden Suite. These dykes are trending E-W with a near vertical dip.

2.3.3 LYKKJEBØ GROUP

The rocks of the metamorphic supracrustals of the Lykkjebø Group include quartzites, feldspathic quartzites, feldspathic mica-schists, micaceous gneisses, dolomite and calcite marbles, calc-silicate rocks and various gneisses with much garnet and either kyanite or sillimanite (Bryhni 1958, Bryhni et. al. 1981). Slices of "anorthosite-kindred" rocks from the Eikefjord Group are present within the Lykkjebø Group (Bryhni et al. 1981). Rocks of the Lykkjebø Group are reported from the areas between Eikefjord and Gloppen **(Fig. 2.5)** (Bryhni et al. 1981, Bryhni 1989). Mica-schists and meta-psammites that are present in a narrow belt along the southern side of the Standal Fault have been assigned to the Lykkjebø Group by Bryhni & Lutro (1991a, 1991b, 2000a, 2000b).

The age of formation of the Lykkjebø rocks is uncertain. Geochronological dating of the meta-supracrustals of the Lykkjebø Group have not yielded age of origin, but they are assumed to be of Eocambrian age (Bryhni 1989). Similar meta-supracrustals at the Tafjord–Sotaseter area (**Fig. 2.1**) seem to be **1000 Ma** or more, according to Bryhni (1977). These meta-supracrustals were suggested to be Telemark Supergroup analogues or possibly equivalents of the Eocambrian Sparagmites in eastern South Norway.

The Lykkjebø rocks have been structurally and metamorphically reworked several times. The rocks of the Lykkjebø Group are characterised by the same penetrative mylonitisation as the Eikefjord Group, commonly obliterating all older structures. This mylonite fabric has been interpreted as related to the Nordfjord–Sogn Detachment Zone (Andersen & Jamtveit 1990; Johnston et al. 2007b), or to the analogous “principal shear zone” (Wilks & Cuthbert 1994), formed during the Devonian extensional phase.

The Lykkjebø Group were assigned to the Fjordane Complex by Bryhni (1966). In spite of the great difficulties in deciding the mutual tectonostratigraphic position of the Lykkjebø and Eikefjord Groups due to intense interfolding and interslicing (Bryhni et al.1981), Bryhni & Sturt (1985) assigned the Lykkjebø Group to a Lower

Allochthon position, i.e. situated below the Eikefjord Group which they assigned to the Middle Allochthon **(Fig. 2.1, Table 2.1)**. Opposing this view, Andersen & Jamtveit (1990) and Furnes et al. (1990) – in their map-figures – assigned the Lykkjebø Group to the Middle Allochthon, considering the rocks as an integral sedimentary cover sequence to the Middle Allochthonous Eikefjord Group, and thereby saying that the relationship between the groups was the same as for the Middle Allochthonous Dalsfjord Suite with its sedimentary cover, the Høyvik Group **(Fig. 2.4)**. Furnes et al. (1990) displayed a map-figure which ***included*** the Lykkjebø Group rocks in the Høyvik Group. Johnston et al. (2007b) assigned both the Eikefjord and the Lykkjebø Groups to the Middle Allochthon.

2.3.4 HØYDALSFJORDEN COMPLEX

The name "Høydalsfjorden Complex" was introduced by Vetti (1996). The complex is located between Brufjorden/Førdefjorden to the south and Solheimsfjorden/ Eikefjorden to the north (**Fig. 2.4**). In the study area of the present thesis, the Lower Palaeozoic rocks that form the substrate to the Håsteinen Devonian Massif, are assigned to this complex. The unit will be more comprehensively discussed in Ch. 4 of the present thesis.

In the litterature, very little information is available from the area constituting the Høydalsfjorden Complex, and some general information from the work of the present author will therefore be given: The rocks essentially consist of meta-greywacke (which are frequently of greenish colour) grading into meta-semipelite and locally meta-psammite (Vetti 1988). The area also contains chlorite- and epidote-rich rocks with a strong green colour, and these rocks are interpreted as meta-volcanoclastics (a view also held by Harald Furnes, pers. com). On the islands of Stavøya (between Brufjorden and Høydalsfjorden), and **Alvora** (in Høydalsfjorden) (**Fig. 2.4**), the meta-greywackes and meta-semipelites show primary gradual transitions into large volumes of white-coloured meta-psammites (author's observations). Similar meta-psammitic rocks further north, at Sandviken, have by Osland (1970) been classified as meta-arkoses. Numerous meta-gabbro bodies with intrusive contacts towards the meta-sediment are present, particularly at the neck of land separating the inner parts of the two fjords Eikefjorden and Høydalsfjorden. Minor occurrences of carbonate and black shales are found locally (this work), and small bodies of ultramafics are present at Stavang (**Fig. 2.4**) (Kildal 1970).

Furnes et al. (1990) interpreted the rocks – which are here assigned to the Høydalsfjorden Complex – as part of the sedimentary cover sequence to the Solund–Stavfjorden Ophiolite Complex (S–SOC), and assigned the rocks to their Heggøy Formation **(Fig 2.4)**. The S–SOC is dated at **443 +/- 3 Ma** by U/Pb on zircons from the island of Tviberg (Dunning & Pedersen 1988). The obduction of the rocks probably occurred at the onset of the Scandian continent–continent collision at mid-Silurian (mid Wenlock) times (Andersen et al. 1990). Later workers have set the obduction to **~425 Ma** (e.g. Milnes et al 1997; Fossen & Dunlap 1998), or more generally (Andersen 1998) to Wenlock times (**428-423 Ma** according to timescale of Gradstein & Ogg 1996, and Gradstein et al. 2004). Lundmark & Corfu (2007) suggested that thrusting started prior to **427 +/- 1 Ma**, and this age was based on U/Pb dating of the Årdal Dyke Complex of innermost Sognefjorden, arguing that the dykes were formed from Baltic sediments that were melted below the overriding Jotun Nappe and thereafter intruded into the nappe. The

Høydalsfjorden rocks have been subjected to polyphasal contractional deformation, where the metamorphism was prograde into lower greenschist facies, and where the main foliation is folded in WNW-ESE trending folds which are cut by the sub-Devonian unconformity (Vetti 1988, 1989). Also Jonhston et al. (2007b) reported that the rocks of the unit primarily exhibit greenschist facies assemblages. However, a thermobarometric investigation of one single locality within the unit yielded significantly higher P/T estimates. The locality, situated on the island of Stavøya in the westernmost part of the unit, yielded values of **9.1 +/- 1.3 kbar, 442 +/- 71 °C**, which the authors assigned to the greenschist–blueschist transitiun. Both the northern and southern boundaries of the Høydalsfjorden Complex are defined by faults, the Sunnar Fault and the Standal Fault respectively (**Fig 2.6**). The block of Høydalsfjorden rocks appears to have been dropped down along these faults, juxtaposing the Høydalsfjorden rocks with the tectonostratigraphically underlying units. The Høydalsfjorden rocks have been assigned to an Upper Allochthon position (Bryhni & Sturt 1985). The rocks treated here have previously been named the Sunnarvik Group (Bryhni & Lutro 1991a, 1991b, 200a, 2000b, Johnston et al. 2007b), but the documented presence of sediments and intrusives justifies the renaming of the unit into a “complex” (Nystuen 1989).

2.3.5 HÅSTEINEN DEVONIAN MASSIF

The Håsteinen Devonian Massif (HDM) is located between Førdefjorden and Eikefjorden at the eastern end of Høydalsfjorden **(Fig 2.4)**. The massif represents the highest tectonostratigraphic/stratigraphic level in the area. Apart from the essentially petrographic descriptions of Kolderup (1925), the geology of the Håsteinen Devonian Massif has been essentially unknown (Steel et al. 1985; Roberts 1983) until the accounts of Vetti (1988, 1989, 1996, 1997; Vetti & Miles 1997), and the present thesis. Osmundsen & Andersen (2001) presented some data from the HDM, that were taken from the works by Vetti (op.cit.). Steel et al. (1985) (referring to Kolderup 1925) reported that the sedimentary rocks in the basin were mainly conglomerates with a probable thickness of more than 1 km. Vetti (1988, 1989, 1997b) briefly reported on sedimentary facies, depositional environments and transporting agents, and also subdivided the sediments into formations. Inliers of substrate rocks within the HDM have been noted by Kildal (1970), and Bryhni & Lutro (1991a, 2000a), and models to explain these inliers were briefly discussed by Vetti (1988, 1997b). The age of deposition is considered to be Middle Devonian by correlation with the other Devonian massifs. The western contact was reported as a depositional unconformity by Steel et al.(1985), whereas the nature of the other contacts were reported unknown. The folding of the massif was noted by e.g. Helland (1880); Vogt (1928); Roberts (1983); Torsvik et al. (1987); Seranne (1988); Vetti (1988, 1989, 1996, 1997); Vetti & Milnes 1997; and Osmundsen & Andersen (2001). Eastward thrusting of the HDM has been suggsted by Roberts (1983), referring to Kolderup (1925), from the southeastern margin of the massif.

Since very little information is available from the HDM in the literature, some general information from the author's work is presented: Nearly the whole massif is made up of coarse conglomerates and breccias. Very small units of sandstones are present immediately above the lower contact in the far west and northwest. The contact towards the substrate (i.e the underlying rock complexes), is a depositional unconformity, locally tectonically modified. The bedding delineates an open synclinal fold with a WNW-ESE oriented axial trace and fold axes plunging towards the ESE. The northern fold limb dips towards the SE, and the southern limb towards the NE. Parasitic folds with an axial planar cleavage are present only in a small sandstone unit in the far west. Along the axial trace, the eastward dip of bedding creates a very large bedding-normal cumulative stratigraphic thickness, and the eastward-dipping layers (along the axial trace) rest on a Devonian palaeosurface envelope, that is subhorizontal. Faults affecting the HDM appear to have only negligible displacements.

2.4 STRUCTURAL INVESTIGATIONS BETWEEN FØRDEFJORDEN AND NORDDALSFJORDEN / INNER NORDFJORD

2.4.1 STRUCTURAL DEVELOPMENT OF THE EIKEFJORD AND LYKKJEBØ GROUPS

In the Lykkjebø Group rocks of the Grøneheia area (**Fig. 2.5**) just to the north of the village of Eikefjord, Bryhni (1962) recognised three phases of deformation, where the F_2-folds and the L_2-stretching lineations of the second phase were oriented WNW-ESE. The structures of all three phases were considered to be of Caledonian age (Bryhni 1962).

In the area between Hyen and lake Breimsvatn (**Fig. 2.5**) – which contains rocks of the Fjordane Complex (the later Eikefjord/Lykkjebø Groups) in the west, and basal gneisses of the Jostedalen Complex in the east – Bryhni & Grimstad (1970) described regional-scale folding (F_2) with roughly E-W trending and mutually parallel axial traces, and W-E to WNW-ESE plunging local F_2-folds and L_2-stretching lineations. Bryhni (1977, 1989) showed that these folds could be recognised over an area covering at least **5 km** in the N-S direction and **10 km** in the E-W direction (**Fig 2.5**), and that the parallel orientation of the F_2 and L_2 structural elements was a regional phenomenon. These fold traces are also shown on maps of the Norwegian Geological Survey (NORDFJORDEID: Bryhni 1974, 2000a; HORNINDAL: Bryhni 1972, 2000b; EIKEFJORD: Bryhni & Lutro 1990, 1991a, 2000a; NAUSTDAL: Bryhni & Lutro 1989, 1991b, 2000b; FIMLANDSGREND: Bryhni & Lutro 1994, 2000c). Bryhni (1989) suggested that the orientation of the structural elements was due to Caledonian reactivation of Precambrian structures with the same direction, since the same orientations were present in the Lower Palaeozoic rocks in the outer part of the Sunnfjord region. A map in Krabbendam & Dewey (1998) included the fold traces, and the folds were here interpreted as a result of bulk constriction occurring in a transtensional regime related to sinistral movements along the Møre–Trøndelag Fault Complex. The same folding was also reported by Young et al (2007), who studied structural development and eclogite exhumation in the areas east and northeast of the Hornelen Massif. The authors proposed the name Sandane Shear Zone (SSZ) for the local segment of the Nordfjord–Sogn Detachment Zone. Johnston et al (2007b) also appreciated this folding, in their study of the structural geology, thermobarometry and geochronology of the Eikefjord and Lykkjebø Groups in the area between Eikefjord and Gloppen. The two groups constitute the NSDZ in this area, and the authors introduced the name “Hornelen segment” for this part of the NSDZ. Because the folding effected the Hornelen sediments, the Hornelen detachment, the 39Ar/40Ar ages and K-feldspar age contours in the WGR to the north, the authors suggested that at least some of the folding occurred in the upper crust after **380 Ma**, and (with reference to Root at al. 2005) possibly as late as **335 Ma**. Johnston et al (2007b) furthermore introduced the name “Blåfjellet Detachment” for the brittle–ductile detachment fault located within the NSDZ at a position east of lake Endestadvatnet, roughly **10 km** east of the village of Eikefjord – a fault first shown on the map by Bryhni & Lutro (1991b, 2000b). Johnston et al. (2007b) interpreted the gentle scoop shape of this fault to be a result of the same regional folding.

A number of papers give information on kinematic indicators within the mylonites of the Eikefjord Group, the Lykkjebø Group and the Jostedalen Complex (e.g. Seguret et al. 1989; Chauvet & Séranne 1989, 1994; Séranne et al 1989; Andersen & Jamtveit 1990; Wilks & Cuthbert 1994; Young et al. 2007; Johnston et al. 2007b). The sense of shear is generally described as being top-to-the-west, and the fabric has been interpreted as part of the Nordfjord–Sogn Detachment Zone that formed during the Devonian extension of the Caledonian orogen (Andersen & Jamtveit 1990; Chauvet & Séranne 1994; Wilks & Cuthbert 1994; Johnston et al. 2007b) (see Sect. 2.8).

Torsvik et al. (1987) carried out tectonomagnetic investigations in the Eikefjord/Lykkjebø Groups, the Høydalsfjorden Complex, and the Håsteinen Devonian Massif, and reported WNW-ESE striking and steeply dipping magnetic foliation planes. They concluded that the Devonian deformation of the Old Red Sandstone massifs and related substrates was due to a regional compressive N-S shortening event related to their "Solundian/Svalbardian Orogeny" (Torsvik et al. 1987) (see Sect. 2.8).

2.4.2 COMMENT: INDICATIONS THAT A "MIDDLE PLATE" DOES NOT EXIST

The term "Middle Plate" has been used by some authors to denote the Eikefjord-/Lykkjebø Group rocks in the Eikefjord–Gloppen area (Andersen & Jamtveit 1990; Dewey et al. 1993; Andersen et al. 1994; Berry et al. 1995; Andersen 1998). The idea has been that the rocks situated between the Håsteinen and Hornelen Devonian massifs form a separate extra "Middle Plate" located between the well established Lower and Upper Plates of western Norway. In this "Middle Plate" model, the Nordfjord–Sogn Detachment Zone is placed *below* the "Middle Plate", i.e. appearing along the easternmost boundary of the Eikefjord-/Lykkjebø Group area. The so-called "Hornelen Detachment" is thought to mark the *upper* limit of the "Middle Plate", defining the present fault contact towards the Hornelen Devonian Massif. As mentioned above, Wilks & Cuthbert (1994), who carried out detailed analyses of the Eikefjord and Lykkjebø Group mylonites, gave no support to the "Middle Plate" concept, but viewed the whole area as an extensional top-to-the-west shear zone. The similar view was held by Johnston et al. (2007b). The present author has carried out reconnaissance work in the Eikefjord–Hyen area, and has observed the same extensive mylonitisation of the whole area. This indicates that the area is part of a top-to-the-west shear zone, and no evidence appears to support the presence of a "Middle Plate".

2.4.3 RADIOMETRIC DATING

$^{40}Ar/^{39}Ar$ dating has been performed on the Lykkjebø and Eikefjord Group rocks of the Eikefjord–Gloppen area, i.e. the area situated between Håsteinen and Hornelen. Chauvet & Dallmeyer (1992) dated muscovites at two localities on the north side of the Eikefjord Fault, obtaining plateau ages of **399.2±0.7 Ma** east of Haukå (south of lake Haukåvatnet) (**Fig. 2.4**) vand **402.9±1.1 Ma** near Lykkjebøen (5 km north of Storebru) (**Fig. 2.5**). Subsequently, Berry et al. (1995 - *an abstract without a map and without dating-documentation*) obtained two groups of muscovite ages (*claimed to be plateau ages*); an old pair of **416** and **415 Ma** to the north of the Eikefjord Fault, and a young group of **400**, **399** and **396 Ma** to the south of the fault. The authors interpreted this to indicate differential uplift of the two areas, and suggested that the Eikefjord Fault be a "detachment" within the "Middle Plate", and that movements along this detachment took place at the youngest date of **396 Ma** or later (Berry et al. 1995). However, their suggestion that the area *north* of the Eikefjord Fault contains "older" ages than the area *south* of the fault, is contradicted by the above mentioned "young" ages of **399 and 402 Ma** that Chauvet & Dallmeyer (1992) found *north* of the fault, a paradox which Berry et al (1995) do not comment on, and which casts doubt on the models of differential uplift, and on the brittle and steeply dipping Eikefjord Fault being at all a detachment.

Andersen (1998) presented $^{40}Ar/^{39}Ar$ ages from the Eikefjord–Gloppen area, ages that were referred to as being the same 'white mica' ages as those of Berry et al. (1993, 1995) (i.e. the ages discussed in the above paragraph). Andersen (1998) presented a small-scale regional *map* that showed, for the first time, the *locations* of the dated samples, but still the ages were presented without any dating-documentation. The ages were asserted to be plateau ages. Some of the ages of Andersen (1998) were slightly different from those of Berry et al. (1995). Andersen (1998) presented the following ages: To the *north* of the Eikefjord Fault, the same two ages of **416** and **415 Ma** were reported (**416 Ma** at the Storebru–Hyen road at the shore of lake Krokstadvatnet; 415 Ma at the Storebru–Hyen road north of the northern end of lake Emhjellevatnet), but also an additional age of 396 Ma (from the Hyen–Sandane road at the eastern shore of central part of Hyenfjorden) were included. To the *south* of the Eikefjord Fault, three ages were reported: 404, 399 and 399 Ma. (404 Ma: from the Storebru–Hyen road near the south end of lake Endestadvatnet; 399 and 399 Ma: two closely situated sample localities, apparently from the old Storebru–Naustdal road at the Ramsdalsheia area). As referred above, the original ages of Berry et al. (1995) south of the Eikefjord Fault were **400**, **399** and **396 Ma**, i.e. the ages reported from the area south of the Eikefjord Fault are fairly similar in Berry et al. (1995: **400**, **399** and **396 Ma**) and Andersen (1998: **404**, **399** and **399 Ma**).

Johnston et al. (2007b) used the Sm/Nd method to date garnets from garnet-muscovite schists of the Lykkjebø Group. Samples were collected from three localities: **i)** At Standal, where the Standal fault reaches the sea: garnet core: **422.3 +/- 1.6 Ma**, garnet rim: **407.6 +/-1.3 Ma**. **ii)** Near Svarthumle, **10 km** east of the village of Eikefjord: garnet core: **425.1 +/- 1.6 Ma**, garnet rim: **415.0 +/-2.3 Ma**. **iii)** In Austre Hydalen, **10 km** SSE of Hyen: garnet core: no data, garnet rim: **414.4 +/- 1.6 Ma**.

The garnet core ages of **425–422 Ma** were interpreted to represent the time of *initial* garnet growth during peak upper amphibolite facies metamorphism, whereas the rim ages of **415–407 Ma** were interpreted to represent the *end* of garnet growth during the same peak metamorphism. The ages were seen as dating the metamorphism that existed *prior* to the formation of the NSDZ. The authors noted that "because two-point isochrones cannot test for original homogeneity in $^{143}Nd/^{144}Nd$ among phases or ensure that all phases remained closed to Sm/Nd diffusion, these ages must be interpreted cautiously". However, the authors were of the opinion that "the high closure temperature of the Sm/Nd diffusion, and the similarities in age of the three analysed samples suggest that a maximum age of **425–422**

Ma and a rim age of **415–407 Ma** represent robust ages for the onset and end of [peak] amphibolite facies metamorphism in the Lykkjebø Group". Thermobarometric studies indicated that this prograde, peak upper amphibolite facies metamorphism took place at at **13–18 kbar, 537–618 °C**.

2.4.4 STRUCTURAL DEVELOPMENT OF THE HØYDALSFJORDEN COMPLEX

In the literature, very little information exists on the *structural* development of the Høydalsfjorden Complex, and the structural measurements on the geological maps of the Norwegian Geological Survey (Norges geologiske undersøkelse, NGU) thus represent some of the few data presently available. These maps show a WNW-ESE trend for the structural elements in the whole area between Førdefjorden/Brufjorden and Eikefjorden/Solheimsfjorden (Kildal 1970, Bryhni & Lutro 1991a, 1991b, 2000a, 2000b) (**Fig. 2.4**). A few observations of top-to-the-east shear in the Høydalsfjorden Complex have been reported (Chauvet & Séranne 1989; Chauvet & Brunel 1989; Seguret et al. 1989), and this were interpreted as related to Scandian thrusting of the Complex (Seguret et al. 1989).

2.5 TECTONOSTRATIGRAPHY BETWEEN FØRDEFJORDEN AND DALSFJORDEN

2.5.1 GENERAL

The following section gives a short review of the tectonostratigraphy of the areas between Dalsfjorden/Vilnesfjorden and Førdefjorden/Stavfjorden, as rocks in these areas have been correlated with assumed analogues in the study area of the present thesis. **Fig. 2.4** gives an overview of the geographical location of the units in the area, and **Table 2.2** gives an overview of their tectonic positions. The units are named according to Andersen et al. (1990) and Osmundsen & Andersen (1994). Kolderup (1928) was the first to suggest a tectonostratigraphy for the area and concluded that the sequence was lying right way up. The western part of the area is presented on the 1:50.000 geological NGU maps "Melvær" (Skjerlie 1985a) and "Askvoll" (Skjerlie 1985b) as well as in the accompanying map description (Skjerlie 1984). Note, however, that unit names; assignments of rock sequenses into units; and structural interpretations, of these publications, may differ from the ones used in the present thesis. Udated versions of the 1:50.000 geological NGU maps “Askvoll” (Andersen & Osmundsen 2001) and the eastward neighbouring map sheet “Dale” (Ragnhildstveit et al. 2001) exists as preliminary editions at the the Norwegian Geological Survey (Norges geologiske undersøkelse, NGU).

2.5.2 WESTERN GNEISS REGION (UNIT 9)

Poorly differentiated gneisses define the lowest tectonostratigraphic level in the area. The areal extent of the gneisses are here defined according to map-figures in some publications (Andersen et al. 1990; Furnes et al. 1990; Osmundsen & Andersen 1994), and according to a combination of information and map-figures in others (Swensson & Andersen 1991; Andersen et al 1994; Skår et al. 1994), (**Fig. 2.4**). According to these publications, the WGR are present to the south of an E-W oriented boundary passing through the village of Askvoll, and the gneisses continue further southwards across Dalsfjorden to Åfjorden, as well as eastwards. This areal definition has been used in most of the later papers that have contained map-figures of the area (e.g. Dewey et al. 1993; Andersen 1998; Engvik et al. 2000; Braathen et al. 2004). Radiometric dating (U/Pb zircon) by Skår et al. (1994) gave an age of **1640.5 ± 2.3 Ma** for the mylonitic Askvoll Group (unit 8, see Sect. 2.5.3 below) which is situated north of – and structurally on top of – these Western Gneiss Region rocks, suggesting that the Askvoll Group rocks are Proterozoic (Skår et al. 1994). Consequently, some authors have included the Askvoll group in the Western Gneiss Region (Skår 1996; Andersen et al. 1998; Skår & Pedersen 2003).

In some earlier publications, (e.g. Skjerlie 1969; Furnes & Skjerlie 1972), migmatitic gneisses of the basal type were reported to be present only in a narrow strip midway between Dalsfjorden and Åfjorden. It may also be noted that Bryhni (1966) originally assigned the whole sequence below the Kvamshesten Devonian Massif (i.e. including, for example, the Dalsfjord Nappe with the Dalsfjord Suite, the Høyvik Group, and the Solund–Stavfjorden Ophiolite Complex (**Fig. 2.4**), see below), to the "original" Fjordane Complex (which was part of the "original" Western Gneiss Region of Bryhni et al. 1981), also including the whole area between Dalsfjorden and Åfjorden. The original Jostedalen Complex of Bryhni (1966) was situated east of a curved line drawn from Høyangsfjorden (**Fig. 2.2**) (branch of Sognefjorden) and through the small town of Førde, i.e. located far east of the area in question here.

The use of the terms "Fjordane Complex" and "Jostedalen Complex" has by later workers been essentially abandoned for these units and replaced with "Western Gneiss Region" (or "Complex"). An exception from this is Swensson & Andersen (1990) who used the term "Fjordane Complex" for parts of the "proper" gneisses located from the village of Askvoll and southwards across Dalsfjorden (**Fig. 2.3**), and the term "Jostedalen Comlex" for the areas midway between Dalsfjorden and Åfjorden. The reason for this entirely new use of the terms "Fjordane" and "Jostedalen Complex" was not clearly stated in the paper, but apparently the authors have simply re-interpreted the **1:250.000** map sheet "Måløy" (Kildal 1970) and renamed units in such a way that the unit "Basalgneis-komplekset" apparently was named "Jostedalen Complex" and the unit "Metamorfe superkrustaler" (assumed late-Prekambrisk to Cambro-Siluian age) was named "Fjordane Complex". Skår et al. (1994) used the terms "Fjordane" and "Jostedalen Complex" in the same way.

Numerous eclogite-bodies occur in the WGR rocks on both sides of Dalsfjorden, and these developed during the subduction of the Western Gneiss Region (i.e. Baltica) below the Laurentian craton during the Silurian–Devonian Scandian phase of the Caledonian orogeny (Andersen et al 1994). The eclogites have been particularly well studied at two localities: Firstly, at the small peninsula of Vårdalsneset on the north side of Dalsfjorden, where the structural development of the eclogites was studied by Andersen et al (1994), and where the P-T relationships were investigated by Engvik and Andersen (2000), who obtained a **T = 680 +/- 20ºC** and **P = 16 +/- 2 Kbar** related to the subduction of the continental crust. This temperature complies with the temperature of **c. 600ºC** which has been found in eclogites throughout the Sunnfjord area (Griffin et al. 1985). Secondly, at the small island of Bårdsholmen near the south fjord-side, where Engvik et al. (2000) studied the structural, mineralogical and petrophysical devlopment of the eclogites and neighbouring rocks.

2.5.3 ASKVOLL GROUP (UNIT 8)

The group is geographically present as a strip of land between the Dalsfjord Nappe in the north and the fjord Dalsfjorden in the south (with exception of two small areas: the little peninsula south of the village of Askvoll, and the southern part of the island of Atløy). The extent of the Askvoll Group is here defined according to Skår et al. (1994).

The term "Askvoll Group" was introduced by "Skjerlie (1969), originally including the rocks on both sides of Dalsfjorden. He interpreted these rocks to be a 5000 metres thick sequence of meta-supracrustals overlying the "basal gneisses". Skjerlie (1969, 1974) reported eclogites from the lower parts of his extensive "Askvoll Group". He assigned also the rocks between the Standal Fault and Førdefjorden, to his "Askvoll Group. Skjerlie & Tysseland (1981) introduced the term "Vevring Complex", which covered the rocks at the peninsula just to the south of the village of Askvoll, and further southwards. The Vevring Complex was furthermore grouped as ("basal") gneisses together with the gneisses to the south of Dalsfjorden. Skjerlie & Tysseland (1981) used the term "Vevring Complex" also for the area between the Standal Fault and Førdefjorden (i.e. the area where the place Vevring is actually situated), as well as for the Florø Peninsula. Their "Vevring Complex" is essentially equivalent to the "Fjordane" and "Jostedalen Complex" of Swensson & Andersen (1991) (**Fig. 2.3**). The Askvoll Group and the underlying "basal gneisses" constituted the Lower Tectonic Unit of Brekke & Solberg (1987) (**Table 2.2).**

An internal tectonostratigraphy has been developed for the Askvoll Group (Swensson & Andersen 1991; Skår 1992; Skår et al 1994). A meta-sedimentary sequence is overlain by meta-igneous rocks, which is further overlain by meta-sediments. The group is extensively mylonitized and has been interpreted as part of the Nordfjord–Sogn Detachment Zone (Swensson & Andersen 1991; Hveding 1992; Osmundsen 1996), a zone which along the Kvamshesten segment is called the Kvamshesten Detachment Zone by some authors (Skår et al. 1994; Osmundsen 1996) (see Sect. 2.8). Numerous eclogite bodies are present both in the lower Askvoll Group and in the "Vevring Complex" of Skjerlie & Tysseland (1981), and Sm-Nd datings by Griffin & Brueckner (1980) have yielded ages between **407** and **447 Ma**.

Skår et al. (1994) dated the Askvoll Group by the U/Pb-zircon method, and obtained an age of **1640.5 ± 2.3 Ma**. They concluded that the Askvoll Group rocks are Proterozoic, implying that earlier assumptions in the literature, of a Lower Palaeozoic age, was not confirmed. The age has led authors to include the Askvoll group in the Western Gneiss Region (Skår 1996; Andersen et al. 1998; Skår & Pedersen 2003).

2.5.4 DALSFJORDEN FAULT

The Dalsfjorden Fault (**Fig. 2.6**) is a shallowly west-plunging, scoop-shaped normal fault which separates the above-lying nappe units as well as the Kvamshesten Devonian Massif, from the Askvoll Group. The name “Dalsfjord Nappe” has previously been used for all units above the Dalsfjorden fault. In the east, the fault separates the Kvamshesten Devonian Massif from the subjacent Askvoll Group-equivalents/the Western Gneiss Region, and in the west it separates the Dalsfjord Suite and higher units from the subjacent mylonitic Askvoll Group (Osmundsen 1996; Osmundsen et al. 1998). Braathen et al (2004) gave a detailed description of the various types of fault rocks present in the fault, and also used this to construct a new field-based classification scheme for fault rocks, which have a wider applicability.

Palaeomagnetic dating of cataclasite and breccia that formed along the fault due to late brittle extensional movements, revealed Late Permian (**250 ± 10 Ma**) and Upper Jurassic/Lower Cretaceous ages (**150 ±**

10 Ma) (Torsvik et al. 1992). Uncertainties associated with these palaeomagnetic ages were estimated to **± 10 Ma**. The palaeomagnetic dating were subsequently reanalysed and supplemented with $^{40}Ar/^{39}Ar$ dating of the same breccias (Eide et al. 1997), suggesting that movements took place in three different periodes: the Late Carboniferous (**310-287 Ma**); the Late Permian to Early/Middle Triassic (**253-238 Ma**); and in the Mid Jurassic to Early/LateCretaceous (**162-96 Ma**), i.e. consistent with the palaeomagnetic ages.

These brittle movements overprint the extensive ductile extensional mylonites (related to the Nordfjord–Sogn Detachment Zone) which are penetratively developed in the subjacent Askvoll Group. The late faulting has affected the base of the Kvamshesten Devonian Massif, which for a W-E distance of **13 km** rests directly on the thin zone of fault rocks present above the subjacent detachment mylonites (Osmundsen 1996; Osmundsen et al. 1998). Although the contact below the Kvamshesten Devonian Massif is a fault contact, the primary depositional unconformity between the Devonian sediments and the subjacent rocks has been reported to be locally preserved along the fault (Høysæter 1969; Lindermann 1982). However, this information has not been further discussed by later publications dealing with the Dalsfjorden fault (e.g. Osmundsen 1996; Osmundsen et al. 1998; Osmundsen & Andersen 2001; Braathen et al. 2004).

It was earlier assumed that the present position of the Dalsfjord Nappe, on top of the Askvoll Gp, was entirely due to Caledonian E-ward (or SE-ward) thrusting (Skjerlie 1969, 1974; Torsvik et al. 1986; Brekke & Solberg 1987), and it was also assumed that the nappe was still situated in its original thrust position. In this scenario, mylonites of the Askvoll Group were viewed as having formed ***during*** the thrusting. However, the recent appreciation of the kinematic indicators with consistent top-to-the-west movement, in the mylonites of the subjacent Askvoll Group (interpreted as part of the Nordfjord–Sogn Detachment Zone), and the presence of the brittle low-angle Dalsfjorden Fault, show that the present position of the "nappe" is due to large-scale extensional movements (Norton 1986, 1987; Seranne & Seguret 1987; Andersen & Jamtveit 1990; Swensson & Andersen 1991; Hveding 1992; Torsvik et al 1992; Osmundsen 1996; Eide et al 1997, 1999; Andersen et al.1998; Braathen 2004).

2.5.5 DALSFJORD SUITE (UNIT 7)

2.5.5.1 GENERAL DESCRIPTION

The Dalsfjord Suite was first described in detail by Kolderup (1922), who used the term “mangerite syenites” for the rocks. The suite consists of syenitic to monzonitic meta-intrusives (now largely gneisses) with a high content of meso-perthite, and later intrusions of anorthosite and gabbro (Kolderup 1922; Brekke & Solberg 1987; Andersen et al 1990; Osmundsen 1996; Corfu & Andersen 2002). The rocks were assumed to be of Precambrian age on the basis of lithological correlation with the rocks in the Jotun Nappe (Milnes & Koestler 1985, Milnes et al 1997; Andersen et al. 1998).

Corfu and Andersen (2002) carried out U/Pb dating of rocks in the unit (which they called the Dalsfjord Complex), to test the commonly assumed correlation with the Jotun Nappe rocks. A *monzonite* from the island of Atløy yielded a zircon upper intercept age of **1,634+/-3 Ma**, interpreted as the age of magmatic crystallisation of the protolith; and a lower intercept age of **882+/-29 Ma**, considred to represent Sveconorwegian metamorphic reworking (although the age was considered to be a bit too young). A titanite from the same rock sample yielded the two ages of **>960 and <920 Ma**, interpreted as a two-stage growth history related to the Sveconorwegian event. A *gabbro*, intrusive into the monzonite, was sampled at Stordalsvatn in the eastern part of the unit. The zircon gave an upper intercept age of **1464+/-6 Ma**, viewed as the time of intrusion, and a lower age of **925+/-50 Ma**, once again indicating the age of metamorphism.

The Gothian age of **1634+/-3 Ma** referred above for the Dalsfjord *monzonite* corresponds well with zircon upper intercept ages of **1694+/-20** and **1666+26/-23 Ma** in the Jotun Nappe Complex (Corfu & Andersen 2002, referring to Schärer 1980), as well as with four recent U/Pb zircon ages from the Upper Jotun Nappe, yielding ages of **1660 +/- 2 Ma**, **1634 +/- 4 Ma**, **1634 +/-4 Ma**, and **1630 +/- 30 Ma** (Lundmark at al. 2007). The **1464+/-6 Ma** age referred above for the Dalsfjord *gabbro* corresponds to a phase of magmatism which is widespread throughout Laurentia and Baltica, and which is also present in the Western Gneiss Region and the Jotun Complex (Corfu & Andersen 2002, referring to Åhäll & Conelly 1998; Tucker et al. 1990/Selnes gabbro; Gabrielsen et al. 1979; respectiveley). The fairly extensive Sveconorwegian reworking of the Dalsfjord Suite can also be seen in the Jotun rocks (Corfu & Andersen 2002, referring to Schärer 1980; Lundmark et al. 2007). In the Western Gneiss Region, the Sveconorwegian reworking is absent in the northern part, and modest in the southern part, implying that the WGR rocks differ from the Dalsfjord and Jotun rocks in this respect (Corfu & Andersen 2002). Both the Dalsfjord and Jotun units are also unconformably overlain by Late Precambrian sedimentary cover sequences: the Høyvik Group rests on the Dalsfjorden unit, and Valdres sparagmites on the Jotun rocks (Lundmark et al. 2007). The sediments are viewed as deposited on the Baltic passiv margin. In summary, both the Dalsjord and Jotun units formed in the Gothian orogeny, were reworked in Sveconorwegian times, were covered by sediments of Baltic margin type, and today occupy a corresponding tectonostratigraphic position. It is therefore concluded that the two units may be correlated (Corfu & Andersen 2002; Lundmark et al. 2007).

The base of the Dalsfjord suite is defined by the Dalsfjord Fault, and the upper boundary, as mentioned above, is defined by the Høyvik Group, which is lying non-conformably on top (Andersen & Dæhlin 1986). The rocks in the Dalsfjord Suite have been assigned to the Middle Allochthon and correlated with the Eikefjord Group of the Eikefjord–Gloppen area (**Fig. 2.5**) (Bryhni & Sturt 1985; see also map-figures in: Andersen & Jamtveit 1990; Furnes et al. 1990; Dewey et al 1993; Andersen et al 1994; Andersen 1998).

2.5.5.2 PERALKALINE, ULTRAPOTASSIC SYENITE DYKES

Two peralkaline, ultrapotassic syenite dykes are present in the Dalsfjord Suite to the north of Dalsfjorden, across from the village of Dale (Furnes et al. 1982). Dating of the dykes by the K/Ar whole rock method yielded Middle Permian ages (**256-262 Ma**, Furnes et al. 1982). The analyses and dating have been performed on old samples collected as "sandstone dykes" by Kolderup in 1922, and the precise localities have not been relocated (Furnes et al. 1982). The ages obtained have been interpreted as related to strong hydrothermal alteration, probably related to movement along the Dalsfjord Fault (Furnes et al. 1982; Torsvik et al. 1997), and the age of intrusion/crystallisation is thus unknown. The E-W trend, which is ***orthogonal*** to most Permian/Mesozoic dolerite (subalkaline tholeiitic) dykes elsewhere in Sunnfjord, and different geochemistry, indicate that the dykes north of Dale are not related to the doleritic Sunnfjord dykes (Torsvik et al. 1997). Palaeomagnetic dating of the syenite dykes north of Dale proved unsuccessful (Torsvik et al. 1997).

2.5.6 HØYVIK GROUP (UNIT 6)

The Høyvik Group is present in the area extending from the island of Atløy and eastwards along the peninsula of Staveneset to the lake Markavatn, and can also be found on the Dalsfjord Suite-rocks just east-northeast of Atløy island (**Fig. 2.4**). The group essentially consists of meta-psammitic rocks with local horizons of mica-schist and carbonates, intruded by pre-orogenic mafic dykes. The lower contact towards the Dalsfjord Suite was initially interpreted as a tectonic contact (Skjerlie 1969, 1974), but later investigations have shown it to be a tectonically modified unconformable depositional contact (Andersen & Dæhlin 1986; Dæhlin & Andersen 1988; Brekke & Solberg 1987; Andersen et al. 1990). Together, the Dalsfjord Suite and Høyvik Group constitute a basement-cover pair (Andersen et al. 1998), and the Høyvik sediments are generally assumed to represent a Late Proterozoic (miogeoclinal) cover on the continental margin of Baltica (Andersen et al. 1998; Corfu & Andersen 2002).

Brekke & Solberg (1987) divided the Høyvik Group into four formations, which from bottom to top were: **(1)** Granesund Formation: sub-arkoses with primary structures showing right way up, and with primary unconformable and tectonically modified contacts towards the underlying Dalsfjord Suite. **(2)** Kvitenes Formation: present on the island of Atløy as mica-quartz-schists with abundant quartz lenses and magnetite porphyroblasts, and local occurrences of garnet-mica-schist. **(3)** Atløy Formation: consists of quartzitic to arkosic meta-sandstones. Abundance of meso-perthite fragments led Brekke & Solberg (1987) to the conclusion that the Dalsfjord Suite was the source area for the sediments. **(4)** Laukeland Formation: consists of mica-quartz-schists with a high content of quartz veins. It thus resembles the Kvitenes Formation, but differs from it by the horizons of carbonates, pillow lavas and greenschists. The top of the formation is defined by an angular unconformity towards the Herland Group.

After the pre-orogenic dyke intrusions, the Høyvik Group experienced orogenic polyphase deformation and metamorphism (biotite grade of greenschist facies) followed by uplift, erosion and deposition of the unconformably overlying Herland Group of **Wenlock** age (Brekke & Solberg 1987). (**Wenloch = 428–423 Ma**, Gradstein & Ogg 1996; Gradstein et al 2004). This pre-Herland Group uplift has been dated by the $^{40}Ar/^{39}Ar$ method (Eide et al. 1997; Andersen et al. 1998) on muscovites from the meta-sediments of the Høyvik Group, yielding ages around **447 ± 4 Ma**. (Andersen et al. 1998). Such a pre-Scandian cooling-uplift event most likely indicates that the Dalsfjord–Høyvik basement-cover pair was affected by an orogenic event during the **Late Ordovician (latest Caradoc/earliest Ashgill)** time (Andersen et al. 1998). The dating thus improves the age constraint for the **Late Ordovician** (pre-Scandian) orogenic event, which was previously considered as merely "pre-Wenlock". The subsequent Scandian (**Late Silurian**) contractional orogenic event and the Devonian late-/post-orogenic extensional event resulted in only minimal Ar-loss in the Høyvik Group rocks, so the radiometric clock was not reset to Scandian ages. This implies that the temperature of the Høyvik rocks did not exceed the retention level of **350–400°C** after the rocks were burried below the overriding Scandian nappes, indicating that the thickness of the nappe pile was less than **~12 km** (assuming a thermal gradient of **30°C/km**).

The Høyvik Group has been assigned to the Middle Allochthon together with the Dalsfjord Suite (Bryhni & Sturt 1985; Andersen & Jamtveit 1990; Swensson & Andersen 1991; Andersen et al. 1998). The group has also been correlated with the Lykkjebø Group in the area between Eikefjord and Gloppen (**Fig. 2.4** and **2.5**) (Andersen & Jamtveit 1990; see also regional map-figures in papers of Andersen et al. 1990; Furnes et al. 1990; Skjerlie 1992*;* Andersen et al. 1994; Osmundsen & Andersen 1994; Osmundsen 1996; Andersen 1998; Torsvik et al. 1998; Braathen 1999). The Høyvik sediments have been correlated also with the Valdres Sparagmite, which lies on the Jotun Compex rocks (Andersen et al. 1998).

2.5.7 HERLAND GROUP (UNIT 5)

The Herland Group is located just southeast of Herland on the island of Atløy (**Fig. 2.4**), and is here well exposed at Brurestakken. The lower contact is a well exposed, depositional angular unconformity against the subjacent Høyvik Group. The Herland Group defines a transgressive sequence consisting of conglomerates, meta-sandstones and carbonates, and has been interpreted as continental margin deposits. Solberg (1987) and Brekke & Solberg (1987) defined four formations, which from base to top were: Sjøralden, Skarvika, Brurestakken and Bjørneneset Formations. After reinterpretation, these units were replaced by two formations by Berg (1988), a subdivision also applied by Andersen et al. (1990): **(1)** Sjøralden Formation (containing the previous Sjøralden and Skarvika Formations), and **(2)** Brurestakken Formation (as previously). The meta-sedimentary Bjørneneset Formation was in stead included in the above-lying Sunnfjord Melange.

(1) Sjøralden Formation (positioned at bottom): basal conglomerates deposited on top of the angular unconformity, interpreted as derived from the Høyvik Group (Brekke & Solberg 1987). The conglomerates are succeeded by cross-bedded (meta-)arkoses and (meta-)quartzites. At Herland, the presence of rugose corals give an **Ashgill-Llandovery** age of deposition (Skjerlie 1974) (**Ashgill = 449–443 Ma,** Gradstein & Ogg 1996, **Llandovery = 443–428 Ma**, Gradstein & Ogg 1996; Gradstein et al 2004), but a newer discovery of pentamerides indicates a **Wenlock** age (M. Johnson pers. com. in Andersen et al. 1990) (**Wenlock = 428– 423 Ma**, Gradstein & Ogg 1996, Gradstein et al. 2004). The upper part of the formation contains black shales, and along these, Andersen et al. (1990) have postulated a décollement surface related to the Scandian thrusting of the Solund–Stavfjorden Ophiolite complex. **(2)** Brurestakken Formation (positioned at top): the formation consists of two cyclic units, each starting with conglomerates and meta-sandstones gradually changing upwards to carbonate and carbonate-bearing schists (Brekke & Solberg 1987).

2.5.8 SUNNFJORD MELANGE (UNIT 4)

The Sunnfjord Melange defines a belt going from the island of Tviberg in the west and via Atløy and Staveneset to Markavatn in the east (**Fig. 2.4**) (Carlsen 1989; Andersen et al. 1990; Osmundsen 1990; Alsaker 1991; Alsaker & Furnes 1994). On the island of Atløy, Brekke & Solberg (1987) assigned the melange to their Bjørneneset Formation which was the uppermost formation in their Herland Group. This interpretation was disputed by Berg (1988) and Andersen et al. (1990), and the latter authors re-interpreted the unit as an ophiolitic melange situated between the Middle and Upper Tectonic Unit. (The terminology of "Tectonic Units" was defined by Brekke & Solberg 1987) (**Table 2.2**). The melange has been reported to essentially consist of carbonate-bearing greenschists and conglomerates, and to have had both continental and ophiolitic sources (Skjerlie 1969, 1974; Berg 1988; Andersen et al. 1990; Alsaker 1991; Alsaker & Furnes 1994). Alsaker & Furnes (1994) divided the melange into two units, the lower Øyravatn Unit of predominantly fine grained calcareous quartz-bearing chlorite-muscovite schists, and the upper Markavatn Unit of mainly meta-greywackes. Both units contain blocks of meta-basalts, marble, and meta-arkose, ranging in size from **tens of metres** to **several kms**. Serpentinite blocks (as well as blocks of chert and lenses of debris-flow-deposited conglomerate), however, are restricted to the Markavatn Unit, and the presence of ultramafic material (serpentinite) in this unit is the basis for establishing it as a seperate unit.

At an early stage of the melange development, a marine basin was present between the obducting S–SOC, and the continental margin, and sediments were continuously supplied from both sides. Geochemical investigations by Alsaker & Furnes (1994) have confirmed that the melange was indeed fed from two sources: i) the obducting S–SOC on the oceanward side, and ii) the continental margin meta-sandstones of the (assumed Eocambrian) Høyvik Group on the continental side. The blocks in the melange, whether of mega-size or smaller, were transported into the basin as olistoliths by submarine debris-flows (Alsaker & Furnes 1994). The sediments were subsequently tectonized below the Solund–Stavfjorden Ophiolite Complex during the obduction of the ophiolite onto continental areas, and Alsaker & Furnes (1994) accordingly classified the melange as a strongly tectonized sedimentary obduction melange.

At the Island of Atløy, conglomerates of the melange can be seen to lie with a depositional primary angular unconformity on the subjacent Herland Group, showing that the melange represents a terrane link between the continental margin rocks and the above-lying oceanic rocks (Andersen et al. 1990). Alsaker & Furnes (1994) suggested that the melange formed adjacent to the Laurentian margin, implying later obduction onto this margin. The question of whether the obduction occurred on the Laurentian or Baltic side, is further discussed below (Sect. 2.5.9.2).

2.5.9 SOLUND–STAVFJORDEN OPHIOLITE COMPLEX (S–SOC) (UNIT 3)

2.5.9.1 HISTORY OF RESEARCH

Parts of the Solund–Stavfjorden Ophiolite Complex (S–SOC) are situated at the following places: western islands of the Solund district; the Hyllestad area (possibly); the island of Atløy and other islands to the west of it; the Staveneset peninsula, with an eastward continuation to the lake Markavatn area; and the Svanøy–Askrova group of islands (**Fig 2.4**). Various parts of the ophiolite have been described by a number of authors, and different periods of research saw focus on different topics:

The **earliest period (lasting until the 1930s)** focused on simple petrographic descriptions and field mapping (Kolderup 1928 and references therein).

In the second period (1960s and 1970s) the research focused on field descriptions and the general Caledonian tectonostratigraphic, metamorphic and structural development (Nilsen 1968; Skjerlie 1969, 1974, 1984, 1985a, 1985b; Furnes 1972, 1973, 1974; Furnes & Skjerlie 1972; Gale & Roberts 1974; Gale 1975; Furnes et al. 1976), with the first geochemical studies appearing in Gale & Roberts (1974), Gale (1975) and Furnes et al. (1976).

The third period (1980s) started with the discovery of *ophiolite*s in the Norwegian Caledonides (Sturt & Thon 1978; Sturt et al. 1979) – and the first authors to apply the ophiolite term to the Solund–Stavfjorden rocks were Furnes et al. (1980). This period saw research on a wide range of topics: **(1)** Extensive general geochemical investigations (e.g. Furnes et al. 1982b); **(2)** Correlation and classification/grouping of the Norwegian ophiolites into an old "Group I" of Early Ordovician age, and a young "Group II" of Late Ordovician age (Furnes et al. 1981, 1985; Sturt 1984; Sturt et al. 1984; Sturt & Roberts 1991), with the Solund–Stavfjorden Ophiolite Complex belonging to the Early Ordovician "Group I"; **(3)** Field studies of magma chamber margins (Pedersen 1986); **(4)** Radiometric dating of the S–SOC (Pedersen & Dunning 1986; Dunning & Pedersen 1988), yielding an age of **443 +/- 3 Ma**, (Late Ordovician) for the formation of the Solund–Stavfjorden Ophiolite Complex; **(5)** This dating, together with ages of other ophiolites, led to a revision of the grouping of the Norwegian ophiolites, with the S–SOC now beeing shifted from the Early Ordovician "group I" to the Late Ordovician "group II" (Pedersen et al. 1988). The revision was accompanied by the use of new *names* for the ophiolite groups: the name "Group I ophilites" was replaced by the "old group" (or "Early Ordovician ophiolite complexes"), and "Group II ophiolites" by the "young group" (or "Late Ordovician ophiolite complexes") (Pedersen et al. 1988; Pedersen & Furnes 1991).

- **The fourth period (1990s to present)** is characterized particularly by the focus on magmatic and tectonic processes occurring during ocean floor spreading and formation of the ophiolite at the oceanic stage (tectonomagmatic models) (Skjerlie 1988; Skjerlie et al. 1989; Johansen 1989; Skjerlie & Furnes 1990; Furnes et al. 1992, 1998, 2000, 2001, 2003a, 2003b; Dilek et al. 1997; Fonneland 1997; Ryttvad 1997; Ryttvad et al. 2000; Fonneland-Jørgensen 2005; Furnes et al. 2006). However, several other important aspects are also treated, such as: **(1)** comparison of palaeo-geotectonic environment with present-day analogues (Furnes et al 1990; Dilek et al. 1995); **(2)** the discovery of the obduction-related Sunnfjord Melange below the ophiolite (Andersen et al.1988; Andersen et al. 1990; Alsaker & Furnes 1994); **(3)** models for geotectonic environments (Pedersen & Furnes 1991; Skjerlie 1992); **(4)** models suggesting on which side of Iapetus the ophiolite formed and was obducted, and by what mechanism (Pedersen et al. 1991, 1992; Pedersen & Dunning 1993; Fonneland 2002**;** Fonneland & Pedersen 2002a); **(5)** provenance studies of the turbiditic cover to the ophiolite (Skjerlie 1992; Pedersen & Dunning 1993; Fonneland 2002, 2002a), **(6)** field apparence of parts of the ophiolite pseudo-stratigraphy (Skjerlie & Furnes 1996), and **(7)** hydrothermal evolution at the oceanic stage (Fonneland-Jørgensen et al. 2005).

- **In the fifth period (2000s)**, the focus is on the new research dicipline “bio-alteration”. A decade ago, it was appreciated that bacterias/microbes play an important role in the alteration of basaltic rocks at the present ocean floor (e.g. Thorseth et. al. 1995). Furnes et al. (2002) investigated chilled margins of pillow lavas from the Solund–Stavfjorden Ophiolitie Complex, and found bio-generated textures, even though the rocks had suffered Scandian greenschist-facies metamorphism and deformation.

Generally, the Solund–Stavfjorden Ophiolite Complex, with the above-lying Heggøy Formation (which is the primary oceanic sedimentary cover to the ophiolite, and thus part of the Stavenes Group, see below), developed in a marginal basin (back-arc basin) located between a near-by situated continental margin on one side and an active subduction system with a volcanic island arc on the other side. The present-day Andaman Sea region may provide a realistic model for this geotectonic situation (Furnes et al. 1990, 2003b).

2.5.9.2 OPHIOLITE OBDUCTION ONTO LAURENTIA OR BALTICA: THE CONTROVERCY

In the 1990s, a controversy existed as to whether the Scandinavian “old group” and “young group” ophiolites were initially obducted onto the Laurentian side of Iapetus, or the Baltic side. Sturt & Roberts (1991) argued that *all ophiolites* were accreted onto the Baltic side. Pedersen et al. (1992) disagreed, and concluded that the “old group” ophiolites were emplaced onto the Laurentian side, as documented by the presence of Laurentian fossils and Archaean detrital zircons. The Laurentian affinity of the “old group” has later been confirmed by other studies (Pedersen & Dunning 1997; Fonneland & Pedersen 2002a, in Fonneland 2002).

Regarding the “young group”, i.e. the Solund–Stavfjorden Ophiolite Complex, most publications in the 1990s suggested that also this ophiolite was probably obducted onto the Laurentian continental margin (Pedersen et al. 1992; Pedersen & Dunning 1993; Alsaker & Furnes 1994). This suggestion was based on the finding of inherited zircons of *Archaean* age (U-Pb dating by Pedersen & Dunning 1993) in the cover sediments (the Heggøy Formation of the Stavenes Group) of the Solund–Stavfjorden Ophiolite Complex — the argument beeing that rocks of *Archaean* age (possible source rocks) had so far only been recorded on the Laurentian side

(Pedersen & Dunning 1993). Another argument was that *Early to Mid-Ordovician* zircon and titanite grains were also recorded in the Heggøy cover sediments, and according to Pedersen & Dunning (1993), the “old group”/Early Ordovician ophiolites — which the authors suggested were accreted to the Laurentian margin in mid-Ordovician times — were the most likely source for the Heggøy sediments.

However, in subsequent research, Archaean U-Pb isotope ages were reported from the mountain plateau of Hardangervidda (**Fig. 2.2**) in central southern Norway, by Raghildstveit et al. (1994) (**2503 Ma**) and Birkeland et al. (1997) (**2838 Ma** and **2735 Ma**). The two papers interpreted the ages to show that crust-forming processes took place in southern Norway in Archaean time, and the papers thereby apparently invalidated the "Archaean rock"-argument for formation/obduction of the S–SOC on the Laurentian side.

Recently, Bingen et al. (2001) reported Archaean U-Pb zircon ages from *metasediments* at two areas, (i) Hardangervidda and (ii) the valley Hallingdalen to the NE of Hardangervidda. The (i) Hardangervidda metasediments belong to the Festningsnutan and Hettefjorden Groups. These metasediments were reported to be semi-conformably associated with granitic orthogneisses that were dated to have formed between **1650-1500 Ma**, and dating of the depositional age of the sediments themselves yielded an age younger than **1572+/-10 Ma**, i.e. the rocks were of Gothian age. Detrital zircons from the Festningsnutan Group produced the three zircon ages **2.76**, **2.47**, and **2.24 Ga**, and the Hettefjorden Group yielded nine ages, of which eight ages centred around **2.80–2.75 Ga** and one age was **3.25 Ga**. (ii) The Hallingdalen sediments were quartzites of the Hallingdal Complex sampled near Eggedal, a region being part of the socalled Telemark Sector. Seven Archaean zircon grains were obtained, yielding ages in the intervals **3.13–3.12** and **2.86–2.68 Ga**. The maximum age for deposition of the sandstone was estimated to **1.71 Ga**, based on zircons of magmatic orogin.

Bingen et al. (2001) interpreted the metasediments as having been derived from the Svecofennian domain. Apparently, this presence of Archaean zircons on the Baltic craton would further weaken the argument of obduction of the Heggøy Formation/Solund–Stavfjorden Ophiolite Complex onto the Laurentian craton. However, Fonneland & Pederesen (2002a; in Fonneland 2002) stated that Archaean *basement* is not exposed in southern Norway, and that the nearest Archaean province is located approximately 1000 km to the north, in the northern part in the Svecofennian shield. Fonneland & Pedersen (op.cit.) further stated that the areal extent of the two mentioned Hardangervidda formations are limited, and that they represent an insignificant source for the Archaean zircons in the Cambro–Silurian sequence of the Baltic margin.

Also the history of the Dalsfjord Suite Nappe could potentially be affected by the interpretation of Baltic versus Laurentian obduction-side for the Solund–Stavfjorden Ophiolite Complex. If the ophiolite was obducted onto the Laurentian side, this apparently would imply that also the Dalsfjord Suite Nappe was originally situated on the Laurentian side. This is due to the fact that the obduction-related Sunnfjord melange, which formed below the ophiolite during the ophiolite obduction, rests with a partly unconformable contact on the subjacent Herland Group (Andersen et al. 1990), which again has a clearly unconformable contact against the underlying Høyvik Group, which again lies with an unconformable contact on the Dalsfjord Suite.

The “obduction-side” problem was further adresses by Fonneland & Pedersen (2002a: in Fonneland 2002), who carried out Sm-Nd isotope analyses on three samples from the Heggøy Formation, and obtained Sm-Nd model ages from **1417 Ma** to **1793 Ma**. However, no U/Pb zircon dating was performed on the Heggøy rocks. The authors concluded that the Sm-Nd results, and available provenance data indicated that the Heggøy sediments were deposited in a marginal basin that formed outboard of the Baltic margin, implying that the "young group" Solund–Stavfjorden Ophiolite Complex also formed near Baltica.

Comment: This conclusion of Baltic-side obduction of the S–SOC implies revising the former view that obduction occurred onto the Laurentian side. However, at **443 +/-3 Ma**, when the Solund–Stavfjorden Ophiolite Complex and the cover sedimets of the Heggøy Formation were formed, the Iapetus ocean had narrowed considerably, and the Laurentian and Baltic margins were positioned close to each other (Meyer et al. 2003). Thus, the Heggøy Formation could have been formed close to the Baltic margin, and still received sediments from the Laurentian margin. The full continent-continent collision that occurred soon after, led to subduction of the Baltic margin below the Laurentian margin, and the Heggøy sediments and the ophiolite were then obducted onto the Baltic margin.

2.5.10 STAVENES GROUP (UNIT 2)

The Stavenes Group was proposed as a formal name by Furnes et al. (1990) for the extensive sedimentary and volcanic cover sequence lying conformably on the Solund–Stav-fjorden Ophiolite Complex, and the sequence was subdivided into the following units **(Fig. 2.4)**: **(1)** Heggøy Formation (formal unit): is considered as the primary sedimentary cover to the ophiolite. The formation is present in the Staveneset–Markavatn area; on the islands of Svanøy and Askrova; at Slotteneset headland and Tryggøy island in the Solund area; possibly in the Hyllestad area east of Solund; and was tentatively also assumed to continue in the area between Brufjorden/Førdefjorden and Solheimsfjorden/Eikefjorden, which comprises the Høydalsfjorden Complex of the present thesis. The primary contact towards the subjacent ophiolite may be observed at Staveneset, and also at Slotteneset in Solund. At the important locality Slotteneset, the unit mainly consits of a dark green to black schist, and at the island of Heggøy the unit is dominated by calcareous, green and grey metagreywacke, with local occurrences of intrusive bodies and pillow lava, as well as phyllite, quartzite, mable and volcanoclastics (Furnes et al. 1990). **(2)** Hersvik Unit (informal unit): is located west of the Solund Devonian Massif, and defines the northernmost outcrop of the S–SOC/Stavenes Group-rocks in the Solund area. The present stratigraphic position within – and contact relations towards – the rest of the Stavenes Group are unknown due to the isolated position of the unit. The unit consists of meta-greywacke with internal horizons of volcaniclastics and conglomerates, together with lavas and intrusions. It probably formed close to an evolved, near-continent island arc (Furnes et al. 1990). **(3)** Smelvær Unit (informal unit): is present on the islands of Smelvær and Molvær east of the Staveneset peninsula. Also this unit is isolated, implying that stratigraphic position and contact relations to the rest of the Staveneset Group are unknown. The rocks essentially consist of alkaline basaltic volcanics in the form of pillow lavas, and minor lava flows and volcaniclastics, which are locally intruded by gabbro. The unit presumably formed on/near an oceanic island (Furnes et al. 1990).

On the western part of Staveneset peninsula, the Stavenes Group was subjected to detailed structural analyses (Osmundsen & Andersen 1988, 1994; Osmundsen 1990, 1996), and geochemical investigations (Carlsen 1989). Geochronological analyses were carried out by Skjerlie (1992) who used the Sm-Nd method to investigate the crustal residence time of the constituents of the sedimentary cover sequence (Heggøy formation) to the Solund–Stavfjorden Ophiolite Complex, a method which gives model ages corresponding to the average crustal residence age of the continental source rocks of the sediments. The ages obtained ranged from **1630 to 1870 Ma**, which were interpreted to suggest derivation from Proterozoic and possibly Archaean crust, as the model ages will always give minimum values. Pedersen & Dunning (1993) carried out a provenance study of the cover sequence (Heggøy Formation), based on U-Pb dating of single zircon and titanite grains, and obtained one Archaean age, **2495 Ma**; ten Proterozoic ages, **1959-1024 Ma**; and four Early-Mid Ordovician ages, **492-462 Ma**, i.e. thereby also confirming the ages of the Proterozoic group recorded by Skjerlie (1992). The Precambrian ages were interpreted to indicate that the S–SOC marginal basin formed close enough to a North Atlantic continental shield to receive sediments from the shield rocks; and the early–mid Ordovician ages were interpreted to stem from an Ordovician Caledonian terrane, probably the early Ordovician ophiolite terrane(s) of Western Norway (formed from **c. 500 to 470 Ma**) that was obducted onto such a continent and deeply eroded by Ashgill times (**c. 449 Ma**, time-scale of Gradstein & Ogg 1996). Pedersen et al. (1992) suggested that the cover sequence formed close to the Laurentian margin, arguing that this was the only known source for Archaean zircons. However, as mentioned above, Fonneland & Pedersen (2002a) suggested that the Heggøy formation of the Stavenes Group was deposited near the Baltic margin, implying formation of the Solund–Stavfjorden ophiolite at the same place.

The Kalvåg melange at Bremanger, located west of the Hornelen Devonian Massif (see below), was by Furnes et al. (1990) tentatively suggested to have formed in the same marginal basin as the cover to the S–SOC, but was not included in their Staveneset Group.

2.5.11 KVAMSHESTEN DEVONIAN MASSIF (UNIT 1)

The Kvamshesten Devonian Massif has been described in a number of publications. A general review of the Devonian massifs are given later in Chapter 2 (see Sect. 2.7)

2.6 LOWER PALAEOZOIC ROCKS

2.6.1 GENERAL

The Lower Palaeozoic rocks were briefly described above, in Sect. 2.3 and 2.5, because these rocks were represented in the two tectonostratigraphic successions treated there. The present section (2.6) deals with the Lower Palaeozoic rocks in a broader area, albeit including the areas treated above.

Based on geographical separation, the Lower Palaeozoic rocks in the western part of the area that will be considered in this regional overview (**Fig. 2.4** and **2.6**), can be divided into four regions: **1)** The islands of Bremangerlandet and Frøya in outer Nordfjord (including the islands of Batalden and Skorpa further south) (Sect. 2.6.2), **2)** northern part of outer Sunnfjord (between Førdefjorden/Brufjorden and Eikefjorden/Solheimsfjorden), which includes the Høydalsfjorden Complex of the study area (Sect. 2.6.3), **3)** southern part of outer Sunnfjord (between Førdefjorden and Dalsfjorden) (Sect. 2.6.4), and **4)** the Solund and Hyllestad areas in outer Sogn (Sect. 2.6.5). Rocks of the same type are also found further east, defining a NE-SW trending belt along the western border of the Jotun Nappe Complex (**Fig. 2.2**). Here, the Lower Palaeozoic rocks crop out between the subjacent WGR and the above-lying Jotun Nappe, but these rocks are not further considered here. The rocks of the four regions are shortly described below.

2.6.2 BREMANGERLANDET/FRØYA

The Lower Palaeozoic rocks of the Bremangerlandet/Frøya area **(Fig. 2.4** and **2.6)** may be grouped into two tectonostratigraphic units that are further described below: a tectono-stratigraphically lower "oceanic terrane" (un-named) of metamorphic igneous rocks (Sect. 2.6.2.1); and the sedimentary Kalvåg Melange that lies with a thrust contact on top of the "oceanic terrane". The melange has been intruded by granodioritic and gabbroic plutons (Hartz et al. 1994; Ravnås & Furnes 1995) (Sect. 2.6.2.2), called the Bremanger Granodiorite and the Gåsøy Intrusion respectively (Sect. 2.6.2.3).

2.6.2.1 METAMORPHIC “OCEANIC TERRANE”

The tectonostratigraphically lower "oceanic terrane", which consists of amphibolites, greenstones, gabbros, ultramafic pods and garnet-mica-schists (Hartz et al. 1994; Ravnås & Furnes 1995), may possibly correlate with the Sunnfjord Melange and/or the Solund–Stavfjorden Ophiolite Complex (Ravnås & Furnes 1995). The terrane has experienced higher grade regional metamorphism (Hartz et al. 1994).

2.6.2.2 KALVÅG MELANGE

The Kalvåg Melange (Bryhni & Lyse 1980; Bryhni et al. 1981; Bryhni & Lyse 1985; Ravnås 1991; Hartz 1992; Hartz et al. 1994; Steen 1994; Ravnås & Furnes 1995; Steen & Andresen 1997) contains fragments mainly of arenitic shallow-marine sediments, deep-marine sandy turbidites, bedded cherts, lava-flow basalts and andesites, rhyolites and rhyolitic ignimbrite, (and in addition pebbly mudstones and calcareous rocks), all embedded in a sheared olistostromal "groundmass". The latter comprises mainly debris-flow deposits, locally with interfingering channel-fill conglomerates (mainly consisting of debris of lithic-volcanic arenites and volcanic rocks), and gully-fill turbidites, which both show erosional contacts to the olistostromal groundmass. The melange is classified as a bedded olistostromal sedimentary melange (Ravnås 1991; Ravnås & Furnes 1995), (as opposed to the tectonized obduction-related Sunnfjord Melange, see Sect. 2.5), and was deposited in the lower part of, or at the base of, a fault-controlled, steep submarin slope that experienced persistent mass failure processes. The slope was situated on the side of a volcanic arc, facing a back-arc basin or a marginal sea that had recently formed by back-arc rifting behind an active continental margin. Prior to this back-arc rifting, the Norwegian *Lower Ordovician* ophiolite complexes had already been obducted onto the continental margin and had also been subjected to deep erosion, thus giving two different source-rock associations (or provenance regions): the newly formed rocks of the volcanic arc, and an (?)older ophiolite complex/island arc sequence, i.e. probably the already obducted *Lower Ordovician* ophiolite complexes. The continental margin discussed here has been interpreted as the Laurentian margin of the Iapetus ocean (Bruton & Bockelie 1980; Pedersen et al. 1988, 1991, 1992; Pedersen & Furnes 1991). The setting and evolution of the back-arc depositional basin of the Kalvåg Melange resembles closely that proposed for the Late Ordovician to Early Silurian marginal basins, i.e. the Solund–Stavfjorden Ophiolite Complex (Furnes et al. 1990; Pedersen & Furnes 1991; Pedersen et al. 1991, 1992). The age of the Kalvåg melange is probably **Ordovician–Silurian**, based on findings of monograptides in a meta-pelite (Reusch 1903; Kolderup 1928). The melange was intruded by the Bremanger Granodiorite and the Gåsøy intrusion (Sect. 2.6.2.3), see below.

2.6.2.3 BREMANGER GRANODIORITE AND GÅSØY INTRUSION

The Kalvåg Melange was intruded by two plutonic complexes: the Bremanger Granodiorite, first described by Kolderup (1912), and the Gåsøy Intrusion, first described by Furnes et al. (1989). The Bremanger Granodiorite body was studied by Skjerlie (1992), who maintained this name for the body. In a recent study, Hansen et al. (2002) called the rocks Bremanger Granitoid Complex. The rocks of this complex consists of granodiorites, tonalites and granites, and the Gåsøy Intrusion consists of mainly gabbronorite and diorite (Furnes et al. 1989). Contact metamorphic aureols were developed in the melange, along both plutons. The Gåsøy Intrusion, which Ravnås & Furnes (1995) considered to be the older of the two, yielded a Sm-Nd age of **380 +/- 26 Ma** (Furnes et al. 1989), but subsequent dating by the U/Pb zircon method showed that the intrusion formed at **443 +/- 4 Ma** (Hansen et al. 2002 referring to Pedersen pers. com.). Dating of the Bremanger Granitoid Complex by the same method gave **440 +/- 5 Ma** as the age of formation (Hansen et al. 2002), and hense the two plutonic complexes have overlapping ages. Primary sedimentary structures are preserved in the contact aureole of these intrusions. The Kalvåg Melange experienced regional metamorphism of maximum upper greenschist facies, and was subsequently unconformably overlain by the Hornelen Devonian sediments (Hartz et al. 1994; Hartz & Andresen 1997).

2.6.3 THE AREA BETWEEN EIKEFJORDEN AND FØRDEFJORDEN (HØYDALSFJORDEN COMPLEX)

The northern part of Sunnfjord (between Førdefjorden and Eikefjorden) contains turbiditic grey and greenish meta-greywackes, locally with intrusive meta-gabbro bodies (this work), grading into meta-semipelite and occationally meta-psammite. The rocks have been assigned to the Høydalsfjorden Complex (see Sect. 2.3 and Ch. 4), which encompasses the Lower Palaeozoic rocks of the study area of the present thesis. The Høydalsfjorden Complex rocks have earlier been assigned to the Heggøy Formation of the Stavenes Group, i.e. the sedimentary cover sequence to the Solund–Stavfjorden Ophiolite Complex (Furnes et al. 1990; see also maps in publications of Furnes et al. 1992; Ravnås & Furnes 1995; Skjerlie & Furnes 1996; Furnes et al. 2000). This previous correlation of the Høydalsfjorden rocks with the Stavenes rocks has been based on lithological similarities, short geographical distances, and corresponding tectonostratigraphic positions. Due to lack of conclusive criteria in the Høydalsfjorden Complex, however, it cannot, at the present stage, be ruled out that the Complex is in part associated with the Sunnfjord Melange or the Kalvåg Melange. The Høydalsfjorden Complex incorporates the same rock units that Bryhni & Lutro (1991a, 1991b, 2000a, 2000b) tentatively assigned to their *Sunnarvik Group*, and the latter name was also adopted by Johnston et al. (2007b). However, with sedimentary rocks intruded by gabbros etc., the term “group” should be replaced with “complex” (Nystuen 1989), justifying the name “Høydalsfjorden Complex” suggested in the pesent thesis. These Early Palaeozoic rocks have been assigned to the Upper Allochthon (Bryhni & Sturt 1985), which is the highest tectonostratigraphic level, apart from the Devonian deposits. The Håsteinen Devonian Massif is deposited on the Høydalsfjorden Complex. These Lower Palaeozoic rocks were earlier included in the original Fjordane Complex of Bryhni (1966), as the gneiss region in western Norway at that time was regarded as containing Precambrian, Eocambrian and Cambro-Silurian rocks.

2.6.4 THE AREA BETWEEN FØRDEFJORDEN AND DALSFJORDEN

The southern part of Sunnfjord (between Førdefjorden and Dalsfjorden) consists of the Solund–Stavfjorden Ophiolite Complex which has been dated at **443 +/- 3 Ma**. (Dunning & Pedersen 1988). The various cover sequences to the ophiolite is assigned to the Stavenes Group. The Heggøy Formation of this group contains the primary sedimentary cover to the ophiolite (see Sect. 2.5). Caledonian (Lower Palaeozoic) ophiolitic rocks were also thought to be present in the mylonitic Askvoll Group below the Dalsfjord Nappe (Skjerlie 1969; Furnes et al. 1976; Swensson & Andersen 1991), but U-Pb/zircon-dating of the rocks by Skår et al. (1994) yielded an age of **1640.5 +/- 23 Ma**, taken to indicate that these rocks are Precambrian gneisses. The Herland Group on the island of Atløy has been interpreted as a continental marginal sequence of **Wenlock** age (Andersen et al. 1990) that was deposited contemporaneously with, and as a lateral equivalent to, the primary cover sequence of the ophiolite.

2.6.5 SOLUND–HYLLESTAD REGION

2.6.5.1 SOLUND AREA

In the Solund area, ophiolitic rocks assigned to the Solund–Stavfjorden Ophiolite Complex, or its cover sediments, crop out in two belts: **a)** in the west below the Solund Devonian Massif (Furnes 1972, 1973, 1974, Furnes & Skjerlie 1972; Furnes et al. 1976, 1982, Skjerlie et al. 1989; Fonneland 1997; Ryttvad 1997; Furnes et al. 1998; Ryttvad et al. 2000; Furnes et al. 2000, 2001, 2003a, 2003b), and **b)** in the Hyllestad area to the east of the Solund Devonian sediments. The rocks in the western Solund area will not be further described here. Instead, focus will be on the Hyllestad area (see below).

2.6.5.2 HYLLESTAD AREA

Lifjorden and Hyllestad Complexes

Ophiolitic rocks crop out in a belt between Hyllestadfjorden and Sognesjøen/ Sognefjorden (Tillung 1986, 1999; Bøe 1997; Meidell 1998, Hacker et al. 2003). Tillung (1986, 1999) assigned the ophiolitic rocks (with a cover sequence ?) to his Lifjorden Complex, which were shown to lie tectonostratigraphically on top of the Hyllestad Complex, comprising continental margin sediments of assumed Eocambrian age.

Sogneskollen granodiorite

The Lifjorden Complex was intruded by sheets of granodiorite, first described by Kolderup (1912), who called the intrusion Sogneskollen Granodiorite. Tillung (1999) made an extensive study of the tectonometamorphic development of the whole Hyllestad–Lifjorden area southward to Sognesjøen/Sognefjorden, and also included detailed investigations of the intrusion, which in the work was still called the Sogneskollen Granodiorite. The intrusive rocks were also studied by Skjerlie et al. (2000), who named them Sogneskollen Granitic Complex. The modal composition was by Tillung (1999) reported as predominantly granodiorite, locally tending towards a quartz monzonite. Skjerlie et al. (2000) carried out more detailed geochemical analyses, and classified the rocks as "granitic", meaning that the rocks spanned the range from quartz diorite to granite. Hacker et al. (2003) conducted a study in the Hyllestad–Sognesjøen/ Sognefjorden area focusing on the exhumation of the high-pressure rocks beneath the Sound Devonian deposits. Their study included radiometric dating of the Sogneskollen intrusives (which they called the Sogneskollen Granodiorite), and the U/Pb-zircon method yielded an age of **434 +/- 4 Ma**.

Structural development

The Hyllestad area has been strongly mylonitized in a shear zone that shows a structural thickness of several kms, and shear-sense indicators show top-to-the-west sense of movement. This fabric has been interpreted as part of the Nordfjord–Sogn Detachment Zone (Norton 1987). Tillung (1999) interpreted the plutonic rocks to have intruded *before* or *during* this major phase of mylonite formation. In the outboard Lifjorden Complex, the intrusion *cuts* structures that may be related to the obduction of the ophiolite, but the intrusion is itself *deformed* by the penetrative top-to-the-west shear (Tillung 1999). Therefore, the intrusion might have formed at the end of the obduction of the Lifjorden ophiolite. According to Skjerlie et al. (2000), the intrusion took place during obduction of the Solund–Stavfjorden Ophiolite Complex, and these authors argued that the magma formed by melting of sediments that were buried beneath the hot overriding ophiolite complex and the associated island arcs.

In the areas described above, the contacts between the Lower Palaeozoic rocks and the underlying units are of both primary and tectonic types. In the Håsteinen study-area, the Lower Palaeozoic rocks are separated from the anorthosites of the Eikefjord Group, by the Sunnar Fault **(Fig 2.6)**.

2.7 THE DEVONIAN MASSIFS OF WESTERN NORWAY

In Norway, late Silurian-Middle Devonian Old Red Sandstone massifs are located in four areas; **(1)** along the coast of Western Norway between Fensfjorden and Nordfjord, **(2)** in the coastal areas from the northern part of Møre og Romsdal county to the southern part of Sør-Trøndelag county, **(3)** near the town of Røros in the eastern part of south Norway, and **(4)** in the Oslo Region (Sigmond et al. 1984; Steel et al. 1985). The present review focuses on the deposits in western Norway, which from south to north comprise the four Devonian massifs of Solund (including the islands of Ytre Byrknesøyene/ Holmengrå in western part of Fensfjorden **Fig. 2.2**, and the islands Bulandet/Værlandet north of the Solund area), Kvamshesten, Håsteinen, and Hornelen **(Fig 2.3)**.

It is reminded that, in the present thesis, the term "basin" is used to denote the original Devonian deposits as they were in Devonian times, whereas the term "massif" is used for the present-day occurence of Devonain rocks, that are only remnants of the original basin. In most of the literature, however, the term "basin" is used for both these variants of the deposits. Below, when summaries are given of various papers, the term "basin" is used in the same manner as in the respective papers.

2.7.1 HISTORY OF RESEARCH

The first studies of Devonian sedimentary rocks in western Norway were carried out by Naumann (1824), Keilhau (1838), Irgens & Hiortdahl (1864), Helland (1880) and Reusch (1881). Irgens & Hiortdahl (1864) were the first to suggest a Devonian age for the deposits. Later studies of plant fossils confirmed this age (Kolderup 1904, 1915, 1916, 1926; Nathorst 1915; Høeg 1936, 1945); and studies of crossopterygian fish (Kiær 1918; Jarvik 1949) suggested Middle Devonian ages, with exception of the Bulandet–Værlandet area which is probably Early Devonian (Kiær 1918). The first separate petrographical descriptions of each of the basins were presented in the 1920s, covering Bulandet–Værlandet (Kolderup 1916), Kvamshesten (Kolderup 1923), Håsteinen (Kolderup 1925), Solund (Kolderup 1926), Hornelen (Kolderup 1927a), and Byrknesøyene/Holmengrå (Kolderup 1927b). Also the folding of the massifs, as well as local occurrences of basal mylonites that were interpreted as related to displacements of the Devonian rocks, were first recognised at that time (e.g. Vogt 1928, referring to Kolderup, without citing particular publications).

The next phase of research, from the 1960s and onwards, concentrated on sedimentological aspects, and separation of the marginal and axial sedimentary facies were established where relevant. It was also suggested that the massifs represent individual "basins" (Bryhni 1964a, 1964b, 1964c; Skjerlie 1971; Nilsen 1968, 1969). In addition, the large stratigraphical thickness, especially in the Hornelen Massif, was addressed, and the first models to explain the sedimentation and basin development were presented (e.g. Bryhni 1964a, 1964b).

From the 1970s, the sedimentology, and the relationship between sedimentology and syn-sedimentary (basin controlling) tectonics, were extensively studied. In the Solund and adjacent districts such studies were carried out by Nilsen (1968, 1969, 1973), Indrevær & Steel (1975), Bryhni (1976), Indrevær (1980), and Michelsen et al. (1983a, 1983b). The Kvamshesten Devonian Massif was investigated by Skjerlie (1971), Asphaug (1975), and Bryhni & Skjerlie (1975), with Høisæther (1971) describing the supposed eastward thrusting of the massif. In the Hornelen Devonian Massif, investigations were carried out by Bryhni (1964a, 1964b, 1964c, 1978, 1982), Steel et al. (1977), Larsen & Steel (1978), Steel & Aasheim (1978), Steel & Gloppen (1980), Gloppen & Steel (1981), Larsen et al. (1981), Pollard et al. (1982), Nemec et al. (1984) and Steel (1988). The depositional rate of Hornelen was estimated to **2.5 m/1000 yr** (Nilsen & McLaughlin 1985). In addition to all these publication, more general accounts of the Devonian massifs have been presented by Bryhni (1975), Steel (1976), Steel et al. (1978) and Steel et al. (1985).

From the mid 1980s, the four Devonian massifs have received considerable new attention after the development of tectonic models that explain the formation of the massifs, and to a varying extent also their deformation, as a result of crustal-scale extension related to collapse of the Caledonian Orogen (e.g. Hossack 1984). (See Sect. 2.8). Aspects of this research will be presented below. The presentation starts with works of regional character, i.e. works that deal with several or all four massifs, and proceeds with works focusing on Solund, Kvamshesten, Håsteinen and Hornelen, respectively.

2.7.1.1 WORKS OF REGIONAL CHARACTER

• Roberts (1983) reviewed the Devonian tectonic deformation in the Norwegian Caledonides and its regional perspectives. The *formation* of the basins was seen as a result of extensional tectonics formed by a transtensional regime. Subsequent *deformation* of the basins, however, were viewed as a result of a late Caledonian ("Svalbardian") orogenic phase. All basins of western Norway were considered to have been thrusted SE-wards during this phase.

• Seranne (1988) presented a regional study covering all four Devonian massifs.

• Roberts (1988) gave a review of the timing of Silurian to middle Devonian deformation in the Caledonides of Scandinavia, Svalbard and E. Greenland. The Devonian basins of Norway, including the basins of western Norway, were interpreted to have been thrust-emplaced to their present location in the mountain belt. It was also stated that during the thrusting, some of the basins had Scandian-deformed *basement* attached to them.

• Chauvet & Seranne (1994) made a regional study dealing with various aspects of the Devonian extension, and the formation and deformation of all four of the Devonian massifs. From the northern margin of Kvamshesten, east of lake Markavatnet, they reported that Devonian beds were sepatated from each other by angular unconformities.

• Braathen (1999) focused on brittle faults in the Sunnfjord area, including faults from within or along the margin of the Solund, Kvamshesten and Hornelen Devonian Massifs. The study also dealt with the origin of the E-W folds in the Devonian and substrate. The folds were interpreted to have formed *after* the westward extension that formed the Devonian basins, and were explained as a result of **Late Devonian–Carboniferous** regional N-S contraction related to sinistral transpression along the Møre–Trøndelag Fault Complex (MTFC).

• Osmundsen & Andersen (2001) presented a rewiew of important sedimentological and structural data from all four massifs, and interpreted the basins as beeing a response to large-scale transtensional tectonics related to sinistral transcurrent movements along the MTFC. In this model, the formation and folding of the basins were believed to have occurred simultaneously.

• Svendsen & Jamtveit (2001) studied fluids and halogens appearing as inclusions in veins located in the Solund, Kvamshesten and Hornelen massifs. The inclusions were reported to have formed at the diagenetic/metamorphic boundary. Based on the inclusions, the temperatures for the formation of the veins were estimated to: *Solund*: **T>300°C**, and **305–330°C** based on chlorite thermometry; *Kvamshesten* and *Hornelen*: **T>230°C** and **<270°C**, i.e. **250 +/- 20°C**. The corresponding burrial was estimated to: *Solund*: **13.4 +/- 0.6 km**; and *Hornelen* and *Kvamshesten*: **9.1 +/- 1.6 km**. Salinity of the vein inclusion-fluids in Kvamshesten and Hornelen resembled that of seawater, whereas Solund salinity was *higher* than seawater; and the authors speculated if this could mean that seawater was introduced into the sediments at some stage.

• Johnston et al. (2007a) reviewed the structural development of the Nordfjord–Sogn Detatchment Zone (NSDZ), which controlled the formation of the Upper Plate, on which the Devonian basins of West Norway were deposited. The authors focused on a review of data published on pressure, temperature and radiometric ages of the detachment zone and Lower Plate eclogitic gneisses around each of the basins, in an attempt to constrain the role of the NSDZ in the exhumation of the UHP rocks of the Lower Plate. The work displays a cross section extending from the Solund Devonian basin to the UHP rocks at the islands of Sørøyane (**Fig. 2.1**) within the Western Gneiss Region north of Hornelen. Individual cross sections through each of the Devonian basins of Solund, Kvamshesten and Håsteinen/Hornelen are shown. A three-stage model was presented for the extension in western Norway: 1) exhumation of the (U)HP eclogitic crustal rocks from large depth to the base of the crust, where WGR rocks were here juxtaposed with Lower and Middle Allochthon rocks. 2) Formation of a top-W, ductile, W-dipping shear zone (NSDZ) at amphibolite facies conditions at the base of the crust. The zone overprinted tectonostratigraphic contacts, and gradually developed into a greenchist facies zone with time. 3) Formation of top-W, brittle-ductile detachment faults that soled out in the NSDZ, and that dropped the Devonian basins into contact with the NSDZ.

2.7.1.2 SOLUND DEVONIAN MASSIF

• Michelsen et al. (1983a) and Michelsen (1986a) discussed a number of aspects concerning the Solund deposits, including the role of the Nordfjord–Sogn Detachment Zone in the basin formation; the sedimentation; and the deformation and metamorphism.

• Michelsen et al. (1983b, 1986); and Michelsen (1986b) performed a detailed study of various parts of a mega rock-body situated within the Devonian sediments at the Hersvik area, central Solund. The body consited of gabbro and other rocks that were similar to those found in the substrate below the Devonian. The body was reported to lie parallel to bedding, and was interpreted as a landslide, which they named the Hersvik slide-unit. The authors recognised that the slide consisted of two parts.

• Norton (1983) contributed with detailed informations of the rock-bodies of the Hersvik area, and concluded that they were of landslide origin. From Kråkvåg, near the southeastern margin of the Solund deposits, it was reported that volcanics of Devonian origin were present.

• Seranne og Seguret (1987) reported on internal deformation of the basin, focusing on orientation of long-pebble axes, mineral lineations, cleavage, tenson gashes, etc., and also on metamorphism.

• Chauvet & Seranne (1989) and Seranne et al. (1989) used the data of Seranne & Seguret (1987) in a regional analysis of Devonian extensional tectonics in the Nordfjord–Sogn area.

• Oddvar Bøe (1997) described the structural evelution of the Nordfjord–Sogn Shear Zone at the island of Losna in eastern Solund, but also included descriptions of sheared Devonian conglomerate near the subjacent brittle Solund fault.

• Muri (1998) studied the post-Caledonian brittle faults in the Solund area, and made correlations with offshore faults as apparent from seismic surveys.

• Sturt & Braathen (2001; also report 1999) investigated the deformation and metamorphism of the outer Solund area, and described tight ductile folding and related axial planar cleavage, that were reported to have formed during lower greenchist facies conditions. From the steep mountain wall at the south side of the Lifjell Peninsula, east of the island of Losna, Sturt & Braathen reported that three large sheets of basement bodies (substrate rocks) were present within the Devonian deposits. The bodies were reported to have Devonian sediments unconformably on top, and tectonised contacts at the base, against underlying Devonian sediments. The sheets were interpreted as basement-slices brought into place by SE-directed thrusting. The possibility that the sheets could be landslides was not discussed. All the structures mentioned above were interpreted to be a result of orogenic compression related to the *Solundian phase* of the Caledonian Orogeny. The authors found it tempting to suggest that “the deformation and metamorphism of the Devonian sequences of the Scandinavian Caledonides results from ***collision between an island arc, situated near Shetland, and the continental block of Baltica***” (Sturt & Braathen 2001). (Italicised by the present author). As an alternative explanation for the Solundian, the authors, referring to Braathen (1999), suggsted that the deformation was caused by regional sinistral transpression due to sinestral movements along the Møre–Trøndelag Fault Complex.

Regarding the timing of the Solundian orogeny, contradicting informations were given in the paper by Sturt & Braathen (2001): At two places in the paper, the authors wrote that the contraction (related to the arc-continent collision) took place in the **Mid–Late Devonian** (see page 270: paper abstract; and page 283: third paragraph of right coloumn)**.** However, the paper also stated that the contraction occurred in the **Late Devonian–Early Carboniferous** (see page 283: fourth paragraph of right coloumn). No explanations were given for this difference in age estimates.

<u>Comment: A possible explanation for the two different age estimates published on the Solundian deformation. A comment on Sturt & Braathen (2001)</u>: In all earlier publications by Sturt and co-workers on the Solundian (e.g. Torsvik et al. 1988), the contraction was interpreted to have occurred in the time-range **Late Devonian–Early Carboniferous**. Sturt & Braathen (2001), however, found support for their new age estimate of **Mid to Late Devonian**, as well as for their new model of *arc-continent collision*, in the presence of Middle Devonian lavas at Shetland and Orkney. These lavas had by Thirlwall (1981) been interpreted as subduction-related arc volcanics, and this interpretation was applied by Sturt & Braathen (2001). Thirwall (1981) had also suggested that the closure of Iapetus in the region probably occurred as late as Mid-Devonian, and this apparently led Sturt & Braathen (2001) to suggest the **Mid to Late Devonian** age for the Solundian deformation. The alternative model, however, for Solundian deformation, presented by Sturt & Braathen (2001), was *sinistral transpression along the Møre–Trøndelag Fault Complex*, outlined in detail by Braathen (1999). The latter publication suggested that this contraction occurred in the **Late Devonian–Carboniferous**. In summary, it appears that Sturt & Braathen (2001) assigned the Solundian to the "**Mid–Late Devonian**" in their descriptions of the new "arc-continent collision" model (see Sect. 2.8. for a discussion on the "Solundian phase"), and to the **Late Devonian–Early Carboniferous** when the deformation was seen as related to movements along the Møre–Trøndelag Fault Complex

• <u>Hartz et al. (2002)</u> studiet the rock bodies in the Hersvik area, bodies that had formerly been studied by Michelsen & co-workers, and Norton (see above). Hartz et al. (op.cit) detected four bodies that were found to consist of variable amounts of rhyolite, granite, gabbro and metasedimentary rocks. The interpretation given by Michelsen and co-workers — that the bodies are landslides — was confirmed by Hartz et al. (2002). As a group, the bodies were, by Hartz et al. (op. cit.) named the <u>Hersvik slides</u>, but two individual bodies were given the names Hagevatnet landslide and Kvernhusdalen landslide. Hartz et al. (op.cit) performed a U/Pb zircon dating of rhyolitic lavas in the Kvernhusdalen landslide, and obtained an age of **439 +/- 1 Ma**, corresponding to Llandovery of the Early Silurian (time-cale of Gradstein & Ogg 1996; and Gradstein et al. 2004). Previous speculations that the lavas were of Devonian age and related to the basin formation (e.g. Furnes & Lippard 1983), were thus rejected. The Kråkvåg volcanics, which Norton (1983) interpreted to have actually formed during the Devonian, were not described by Hartz et al. (2002).

<u>2.7.1.3 KVAMSHESTEN DEVONIAN MASSIF</u>

• <u>Torsvik at al. (1986)</u> presented palaeomagnetic data as well as magnetic *fabric* data from the Kvamshesten massif. The orientation of the palaeomagnetic vectors were interpreted to comply with a pole position suggesting a Late Devonian / Early Carboniferous magnetisation age for the Devonian sediments as well as for the subjacent mylonites of the Askvoll group. The presence of syn-tectonic magnetisation (possibly some post-tectonic magnetisation) showed that most of the folding of the massif took place whilst the temperature was higher than the temperature required for magnetic remanence acquisition. (This means that the magnetic vectors maintained a parallel orientation during most of the folding. In the final stages of the folding, however, the temperature was low enough to “freez” the vectors, and the folding then gave a slight spread in the vector orientation). The magnetisation age was interpreted to date a claimed eastward ***thrusting*** of the Kvamshesten massif on the subjacent mylonites, inferred to be a result of the late Devonian / Early Carboniferous Solundian/Svalbardian orogenic phase.

• Smethurst (1990) studied the palaeomagnetics of the Kvamshesten and Hornelen basins. A remanence of probable Devonian age was reported, but its age in ralation to late Caledonian folding was considered uncertain.

• Markussen (1994) described the structural geology between the Grunnevatn fault and the Instelv fault, with particular focus on the structural controle on the sedimentation in this area. [Cand. Scient thesis].

• Osmundsen (1996) and Osmundsen et al. (1998) presented the first modern investigation covering the ***entire*** Kvamshesten massif. The studies covered sedimentary aspects (distribution of sedimentary facies, depositional processes, syndepositional faults, etc.), as well as the later deformation of the basin, which involved folding around an E-W axis, faulting, and westward movement of the basin on the subjacent Dalsfjord fault. The intrabasinal unconformities that were reported by Chauvet & Seranne (1994), were closely studied in the field by Osmundsen (1996) and Osmundsen et al. (1998), who found the unconformities to be absent or bearly detectable, and of low significance. Near the northern basin margin, the Kringlefjellet reverse fault was reported to have thrusted a body of substrate rocks on top of Devonian sediments.

• Bakke (1999) described the relationship between synsedimentary faulting in Kvamshesten and the distribution and migration of sedimentary facies units. [Cand. Scient. thesis].

• Osmundsen et al. (2000) presented a detailed study of the sedimentary architecture of the basin, applying sequence stratigraphic concepts. The basin was interpreted to have formed above a ramp-flat extensional fault.

• Sejrup (2004) reported on the sedimentology of fine- to coarse grained grey sandstones in the Kvamshesten basin. [Master thesis].

• Ranes (2005) studied the sedimentology of fine to coarse grained grey-coloured sandsones in the Kvamshesten basin. [Master thesis].

• Hauso (2006) analysed the structural geology of the Berge Fault, a synsedimentary fault in the Kvamshesten basin. [Master thesis].

2.7.1.4 HÅSTEINEN DEVONIAN MASSIF

Only a limited amount of information has previously been presented from this massif.

• Torsvik et al. (1987) presented the first modern contribution from the Håsteinen massif. The paper presented palaeomagnetic data and magnetic *fabric* data from the massif and surrounding areas. The main magnetic remanence was interpreted to have been aquired in the Devonian. The authors interpreted the magnetisation to be a result of the Solundian/Svalbardian orogenic phase. The magnetisation of the folded Devonian sediments was imposed post-tectonically or possbly late syn-tectonically, implying that the folding ocurred whilst the temperature was still higher than the temperature required for palaeomagnetic remanence.

• Vetti (1988, 1989, 1996) and Vetti & Milnes (1997) presented abstracts with overviews of the geological development of the massif.

• Vetti (1997) presented data that covered aspects of the sedimentology (facies, etc), basin architecture (stratigraphic thickness, etc), deformation (folding, faulting, etc), contact relationships, inliers of substrate rocks, and metamorphism.

• Vetti (1999) presented a popular overview of major geological features of the Håsteinen massif.

2.7.1.5 HORNELEN DEVONIAN MASSIF

• Nilsen & McCaughlin (1985) compared the tectonic framework and depositional patterns of the Hornelen basin with the “Ridge” and “Little Sulphus Creek” strike slip basins of California. The authors estmated the rate of deposition in the Hornelen basin to **2.5 mm/yr**. This implies that the deposition of the **25 km** thick bedding-normal stratigraphy of Hornelen took **10 million years**.

• Torsvik et al (1988) presented palaeomagnetic data and magnetic *fabric* data from the massif. The palaeomagnetic data yielded a clearly Devonian pole position (Fig. 18a of their paper), situated well away from the Carboniferous pole position. Yet, the authors advocated that the magnetisation was a result of a Late Devonian to Early Carboniferous orogenic event, which were taken to be consistent with the Solundian/Svalbardian Orogeny. The magnetisation was reported to have been aquired syn-to post-tectonically, showing that most of the folding of the massif took place whilst the temperature was higher than the temperature required for magnetic remanence aquisition. At the eastern margin of the basin, the authors claimed to have found that the basin rested with a primary contact on the substrate, and also reported that “well preserved weathering profiles” were present. However, no documentation were presented to support this conclusion, even though it was a fundamental re-interpretation of all previous works, which had reported a tectonic contact in the east.

• Norton et a. (1990) presented a “discussion” of the paper of Torsvik et al. (1988). An additional “discussion”, written by Michel Seranne, was attached to the discussion authored by Norton et al. (1990). These two “discussions” presented strong objections to several of the statements in Torsvik et al. (1988), but espesially rejected the existence of the claimed unconformable eastern contact of the Hornelen basin. Strong opposition was particularly given by Norton et al. (1990), who had examined this contact at a number of localities. They had found no signs of a primary unconformity, but instead mylonites and fault rocks, consistent with previous works.

• Smethurst (1990) studied the palaeomagnetics of the Hornelen and Kvamshesten basins. A remanence of probable Devonian age was reported, but its age in ralation to late Caledonian folding was considered uncertain.

• Torsvik et al. (1990) gave a “reply” to the “discussion” by Norton et al. (1990) and the attached “discussion” by Michel Seranne. The reply, which was exclusively adressed to Michel Seranne, unfortunately offered no comments to several of his major objections. For example, the objections to the claimed unconformable contact at the eastern margin of the basin was left in silence. In their reply to Seranne, Torsvik et al. (1990) wrote that they were also preparing a reply to Norton et al. (1990).Unfortunately, this reply has never appeared in the journal.

- Cuthbert (1991) investigated the evolution of the Hornelen basin, by carrying out petrological studies of the conglomerate clasts. It was concluded that the clasts came from a source area containing the same rock types as those now present in the allochthons below the basin. No petrographic linkage was found between the clasts and the Western Gneiss Region. K-Ar-muscovite dating of a (i) schist from the the Upper Plate rocks that form the substrate to the basin, and of a (ii) conglomerate clast from the basin, both yielded the same age of **423 +/- 8 Ma**, indicating that temperatures in the basin never reached up to the blocking temperature of **350–400°C** for the K-Ar system, during maximum burrial. A rock-slide of maximum **100 m** lenght was reported to lie in the conglomerates at the very Hornelen mountain peak, which is situated at the easternmost part of the island of Bremangerlandet. The slide was interpreted to indicate proximity to the source area for the Devonian deposits, an interpretation which is consistent with the slide being positioned close to the northen margin.

- Wilks & Cuthbert (1994) investigated the evolution of the Hornelen basin detachment system, and the implications for the style of late orogenic extension in the southern Scandinavian Caledonides. Although the work focused on the mylontitic detachment zone below the basin, a brief review of the Hornelen basin was included. It was noted that a heave-displacement of **1 cm/yr** of the basin floor (Upper Plate) relative to the Lower Plate, would give a total heave of **50 km** in **5 Ma**. The regional extension and associated basin formation were seen as a result of divergent plate movements between the Laurentian and Baltic plates.

- Folkestad (1995) studied aspects of the sedimentology of the Hornelen basin, focusing on the cyclisity of the layers in the basin. [Cand. Scient. thesis].

- Allen & Hovius (1998) studied sediment supply from landslide-dominated catchments (i.e. steep source areas), and the implications for basin-margin fans. From measurements of margin-fan sizes in Hornelen, they calculated the size of the catchment area that had supplied the sediments.

- Odling et al. (1999) described fracture system geometries in Hornelen, as a part of a larger study on implications of fractures for fluid flow in fractured hydrocarbon reservoirs.

- Odling & Larsen (2000) reported on the vein architecture in sandstones (from the Magnhildskaret area), and implications for the palaeostrain history. Two generations of veins were found: The “Early veins” were reported to be white- or green-coloured and consisted of a fine-grained mineral mass of quartz, calcite, chlorite and epidote. The “late veins”, which cross-cut the early veins, were described as fibrous and coarcer grained. These “late veins” mainly consisted of white-coloured fibrous quartz, locally with calcite. Chlorite occationally occurred within quartz crystalls. The authors drew the following conclusions: The *Early veins* were formed during the Early Carbonifeous period. At this time the basin had reached its maximum thickness of **15 km** or more, and the Devonian deposits had already become lithified to a degree equivalent to their present state. Rapid exhumation in the Early Carboniferous generated high fluid pressures and hydraulic fracturing that led to formation of the early veins (and breccias) along pre-existing weakness zones. The direction of tectonic extension was NW-SE. The fluid source was possibly located in faults present in the rocks ***beneath*** the basin, generating fluids with a temperature in excess of **300°C.** The *Late veins* formed during Late Permian–Triassic times. The stressfield had by this time rotated some **60°** from the NW-SE trending Early veins, to now generating WSW-ESE trending Late veins. The Late veins formed in a lower energy environment with lower fluid pressure and fluid availability, and fluid temperatures were around **300°C**.

- Folkestad & Steel (2001) studied the alluvial cyclicity of the Hornelen basin by means of a multiparameter sedimentary analyses, and considered the stratigraphic implications. In the study, the axial sandstones of the Svelgen area were investigated in detail. The development of the **100–200 m** thick cyclic sandstone beds (cyclothemes) were analysed with regard to their A/S ratio (Accomodation space / Sediment supply), which were the main factors controlling the spatial and stratigraphic pattern of the basin infill. Sequence-stratigraphic concepts were applied.

- Anderson & Cross (2001) described the large-scale cycle architecture in the Hornelen basin, by studying a number of nine large-scale stratigraphic cycles, each being **c. 100 m** thick, along the northern basin margin. The study area contained conglomeratic alluvial fans, braidplain sandstones and lake facies siltstone tracts, and allowed recording of systematic expansions and retreats of these various facies. The difference in angle between alluval fan surfaces and the braidplane was estimated to **4°,** and since the braidplane was subhorizontal during deposition, this implied that the fans had a **surface slope of 4° during deposition**. From north to south across the basin, cycles were reported to have equal thicknesses, with exception of the northernmost part where the alluvial fans are somewhat thicker, indicating higher subsidence near the northern margin. The A/S ratio (Accomodation space / Sediment supply) was studied.

- Anderson (2002) studied the sequence-stratigraphy in the high “Accomodation/Sediment supply” (A/S) regime of the continental Hornelen basin. Three cycles at the northern margin were investigated, and these were located in the same area as those studied by Anderson & Cross (2001). The Sediment supply (= S) versus subsidence (= A, accomodation) were found to be in balance, as evidenced by the absence of erosional surfaces, and by the fact that Base-level-rise and Base-level-fall sediments have equal thicknesses within one cycle. High sediment supply were found to keep pace with the tectonic subsidence, creating an aggradational yet cyclic stratigraphic architecture.

- Larsen (2002a, 2002b) studied folding, faulting and veining in the northern part of the Hornelen basin, and interpreted these features to be a result of dextral transpression during westward extrusion (movement) of the basin along the original basin-controlling E-W trending fault. This basin-controlling fault was seen as a precursor fault that was located just north of, and parallel to, the Bortnen fault which is presently defining the northern margin of the deposits. The movement was considered to have taken place in Late Devonian to Early Carboniferous times. The development was outlined as follows: At the starting point of the model, the basin had been formed, and the beds probably dipped shallowly eastward as a result of syndepositional tilting. During continued westward movement of the basin, dextral transpression occurred along the northern fault margin. This caused the beds to be folded to a near vertical orientation near the margin, thus producing the Ålfoten monocline. The folding was assumed to have taken place in the Late Devonian–Early Carboniferous, as suggested by palaeomagnetic work of Torsvik et al. (1988). During late stages of this folding, and in response to the folding, extensional veins and related faults were formed in the monocline area. (Vein abundance decreases towards south, and disappears **4.2 km** south of the Bortnen fault). The veins formed with an orientation broadly orthogonal to the bedding. Conglomerates and sandstones alike were observed to be cut by the veins, and this shows that the veins formed in well lithified sediments (as did the faults). This implies that also the folding affected well lithified sediments, i.e that the folding occurred at relatively deep stratigraphic levels after considerable burrial. Such folding at large depth contradicts Chauvet & Seranne (1994), who claimed, on basis of apparent unconformities, that the folding were syndepositional, i.e. took place in soft sediments. The geometry of the monocline indicates that the basin had a **minimum** lateral displacement of **5 km** along the northern margin. In the *northern* part of the monocline, near the basin margin, subhorizontal vein orientations indicate a subvertical extension that was caused by the NNW-SSE shortening. The broadly N-S trending and steeply dipping veins in the *southern* part of the monocline reflect roughly E-W extension in these parts. This E-W extension is the regional extension direction present further southwards in the Hornelen basin. The *basin substrate north of the Hornelen basin* was found to contain conjugate strike slip faults that formed in response to constrictional N-S shortening and E-W extension, a strain field that in the literature has been reported to be present on a regional scale. In summary, both the dextral transpression ***inside*** the ***Hornelen basin*** (northen part) and the regional constriction ***outside the basin***, were interpreted in terms of westward extrusion of the Hornelen basin.

- Fonneland & Pedersen (2002b) studied the provenance of the Hornelen Devonian Basin and the related implications for the exhumation history of the Western Gneiss Region (WGR). The study focused on the provenance of the *axial sandstones* of the basin. Data were obtained by: **1)** Detrital zircon U/Pb dating, **2)** Sm-Nd isotope systematics, **3)** major element geochemistry, and **4)** trace element geochemistry. According to the study, the Western Gneiss Region situated north of and east of the Hornelen basin consists of **80%** Early Proterozoic rocks, mainly orthogneisses, that formed in the intervall **1600–1700 Ma** during the Gothian Orogeny (**1750–1500 Ma**). Furthermore, the study conveyed that **20%** of the region consists of Middle to Late Protorozoic granitoids that intruded the Gothic rocks at **c. 1000 Ma**, i.e. during the Sveconorwegian Orogeny (**1250–900 Ma**). Results and interpretations were as follows (numbering as above): **1)** The Hornelen zircons were reported to show major peaks at **1600–1700 Ma** and **1100–1200 Ma**, and the clear overlap with the ages of the Western Gneiss Region (WGR) was taken to indicate that the latter was the source for the Hornelen sediments. (A few of the Hornelen zircons were of Archaean age). **2)** The Nd-Sm data from Hornelen was taken to support the interpretation that the Western Gneiss Region was the source area. **4)** The trace element analyses were found to indicate that the Hornelen sandstones also received a significant component from a *mafic* source, which was suggested to be either the Jotun Nappe or the Solund–Stavfjorden Ophiolite Complex. Samples from the Høyvik Group and the Herland Group, situated west of the Kvamshesten Devonian Massif, were analysed in order to test whether Middle Allochthon rocks (including the Dalsfjord Suite, etc.) could be the main source for the Hornelen sandstones, but was found not to be the main source. All factors considered, the Western Gneiss region was interpreted as the main source for the axial sandstones of the Hornelen basin.

The authors emphasised that this interpretation had consequences for the exhumation history of the Western Gneiss Region around Hornelen, since, in order to supply sediments to the Hornelen basin, the Western Gneiss Region had to be positioned at the ground surface at the start of basin formation, i.e. at the beginning of the Middle Devonian, which would correspond to **394 Ma** according to the authors (who applied the time-scale of Tucher et al. 1998:

Middle Devonian **394–382.5 Ma)** [1]. For uplift ages, the authors referred to $^{40}Ar/^{39}Ar$ muscovite plateau ages obtained by Chauvet & Dallmeyer (1992) and Berry et al. (1995) from the NSDZ and the WGR adjacent to the Hornelen deposits, and concluded that the ages clustered in the interval **403–393 Ma**. The authors also stated that similar ages (from the same method) were obtained by Fossen & Dunlap (1998b), from the décollement zone under the Jotun Nappe, the latter representing an eastern border of the WGR. Referring to a cumulative probability density plot of the $^{40}Ar/^{39}Ar$ ages, appearing in Fossen & Dunlap (1998b), the authors noted that the ages showed a *maximum* at around **400 Ma**. Based on these $^{40}Ar/^{39}Ar$ datings, the authors suggested that at **400 Ma,** the Western Gneiss Region was at a temperature of **300–350°C**, or at approximately **12–14 km** depth (assuming a thermal gradient of **25°C/km**). In their calculations of uplift rates, the authors applied this depth of **10–15 km** at **400 Ma** for the WGR, and in addition also applied the assumption that the stratigraphically *upper part* of the Hornelen sediments had an age of **382.5 Ma** (i.e. upper boundary of the Middle Devonian, cf. the time-scale used by the authors, see above), or older. From this, the authors calculated that the WGR had a minimum uplift rate of **10–15 km** during the interval **400–382.5 Ma**, i.e during **17.5 Ma**, corresponding to a minimum exhumation rate of **0.5–0.8 mm/yr**, which was reported to be the same as the rate *prior* to **400 Ma**. The authors emphasised, however, that the deposition clearly started *earlier* than at **382.5 Ma**, i.e earlier than the end of Middle Devonian, since sediments with an apparent stratigraphic thickness of **25 km** had to be deposited within the Middle Devonian time interval. The lowermost part of the stratigraphy hence received WGR-derived sediments at a much earlier time. Accordingly, the authors underlined that the real uplift rate of the WGR was significantly higher. For comparison. the authors reminded about Himalayen uplift rates of **6 mm/yr**, i.e. ten times higher, which in that region had removed as much as **12 km** of overburden in **2–3 m.y**.

***Comment: Indications that the WGR was NOT the source area for the Hornelen sediments: A comment on Fonneland & Pedersen 1992b*.** The above reasoning by Fonneland & Pedersen (1992b) is based on their assumption that the WGR was positioned at **300–350°C / 10–15 km** depth at **400 Ma**. However, along the fjord of Nordfjord in the Western Gneiss Region just north of Hornelen, uplift ages are substantially *lower* (*younger)* than this. Here, **seven** $^{40}Ar/^{39}Ar$ muscovite datings (Berry et al. 1995; Andersen 1998), and **one** K/Ar muscovite dating (Cuthbert 1991), have yielded plateau ages in the interval **390–385 Ma**, implying a mean age of around **387 Ma**. The **8** sample localities are evenly distributed along an E–W distance that corresponds to a large portion of the E–W lenght of the Hornelen massif. In their reference to the ages of Berry et al. (1995), Fonneland & Pedersen did not discuss these young ages north of Hornelen. According to Fossen & Dunlap (1998b; referring to Dodson 1973; Wagner et al. 1977; Snee et al. 1988; Hames & Bowring 1994), the blocking temperature associated with the $^{40}Ar/^{39}Ar$ muscovite method is **350–400°C**, corresponding to a depth of **14–16 km** (when assuming a thermal gradient of **25°C/km**), i.e. with a mean depth of **15 km**. If we now apply this mean depth of **15 km** for the Western Gneiss Region and the mean age of **387 Ma** for this depth, (and set the “Middle Devonian” to span the interval **394–382.5 Ma**, as applied by Fonneland and Pedersen 2002b), the time interval from **387 Ma** (15 km depth) to **382.5**

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[1] This “Middle Devonian” on the time scale of Tucker et al. (1988), (**394–382.5 Ma**) differs only slightly from the corresponding interval on the time-scale of Gradstein et al. 2004 (**398–385 Ma**).

Ma (surface) is only **4.5 Ma**. This will then be the maximum time interval available for bringing the source rocks from **15 km** to the subaerial surface – in order for the sediments to reach deposition within the Middle Devonian times.

As pointed out by Fonneland & Pedersen (2002b), the deposition of gneiss-derived sediments must have started long before **382.5 Ma** which is the end of the Middle Devonian. To illustrate the timing problem, we can for example assume that the deposition started at **387 Ma**, i.e. **5 Ma** before the end of Middle Devonian. However, and as evidenced by the Ar-ages referred above, the WGR north of Hornelen rested at **15 km** depth at this time and earlier, and therefore obviously cannot at the same time have been a source area for the Hornelen sediments. If we instead assume that deposition occurred as late as around **382.5 Ma**, i.e. at the end of the Middle Devonian (according to the time-scale applied by Fonneland & Pedersen 1992b, see above), the mentioned time interval of **4.5 million years** would now be "available". During these **4.5 m.y**., the gneiss rocks would have to be uplifted from **15 km** depth to the surface, which in itself is geologically possible. However, in addition, the entire Hornelen basin would have to be deposited within the same time frame, since the Middle Devonian fossils of the basin are located at the top of the basin stratigraphy. A critical factor here is the time required for the Hornelen sediments to form. Nilsen & McCaughlin (1985), (referred to by Allen & Hovius, 1998), have estmated the rate of deposition in the Hornelen basin to be **2.5 mm/yr**. This implies that the deposition of the **25 km** thick stratigraphy of Hornelen took **10 million years** in itself. With only **4.5 Ma** years available for both gneiss uplift and basin deposition, and probably, in reality, "**0 Ma"** available for uplift, since deposition must have started already when the WGR rested at **15 km** depth, the model of Fonneland & Pedersen (2002), that the WGR sourced the Hornelen basin, appears to be problematic.

This is also indicated by published data on clast petrography obtained at the eastern margin of Hornelen: As mentioned above, Cuthbert (1991) made a detailed study of the clast petrography of the marginal conglomerates of the Hornelen sediments, and found that none of the clasts were derived from the WGR. One of these sample localities was situated at the eastern margin of the Hornelen deposits. Also for this locality, Cuthbert (1991) drew the conclusion that the conglomerate clasts were not derived from the WGR, but that they instead corresponded to rocks types that is now present on the island of Bremangerlandet. These Bremangerlandet rocks constitute remnants of the Caledonian nappe pile, that has been preserved on the Upper Plate (Bremangerlandet) which now forms the floor below the Hornelen deposits. All the ten localities distributed evenly along the northern margin of Hornelen had clasts of this Bremangerlandet type. At the locality that is placed at the *eastern* Hornelen margin, sediment transport has been roughly towards the west. The conglomeratic fans that originated in this area – but that are now situated lower in the stratigraphy, once supplied sand to the axial sandstones studied by Fonneland & Pedersen (1992b). From the investigations of Cuthbert (1991), it appears that the present-day eastern marginal fans do not contain clasts from the WGR. These fans, that are situated at the stratigraphically highest level in the Hornelen succession, were fed from Caledonian nappes in the east. With such nappes exposed in the east, it appears unlikely that the axial sandsones, that are situated lower in the stratigraphy, were fed from the WGR level, as the WGR appear to have been covered by nappes.

Also published data on temperature/uplift indicate that the WGR was not exposed at the time of ORS deposition: Eide et al. (1999) performed alkali feldspar $^{40}Ar/^{39}Ar$ geochronology on samples from the Sunnfjord region, and constructed cooling curves from the data. The data were obtained from the island of Bårdsholmen, located in the WGR to the south of Dalsfjorden (i.e. south of the Kvamshesten Devonian Massif), and showed that the temperature of the WGR rocks were higher than **285°C** as late as at **360 Ma**, i.e. near the end of the Devonian period. With a thermal gradient of **25°C/km**, this temperature would correspond to a depht of **~11 km**, indicating that the WGR was not exposed during the deposition of the Kvamshesten ORS basin during the Middle Devonian. It is likely that the WGR around the Hornelen Devonian Massif was at a similar depth at the time of formation of the Hornelen basin.

From the above considerations, it seems clear that the Western Gneiss Region was not at the surface during deposition of the Hornelen basin. Instead, the source area for the Hornelen sediments must have been gneissic rocks in Caledonian nappe position, i.e gneissic rocks that were present within the Caledonian allochtons that surrounded the Hornelen basin during deposition, and that have later been eroded away. In the search for a source area for the Hornelen sandstones, it may be useful to first consider the conglomerates of the basin. As referred above, Cuthbert (1991) investigated clast petrography in the alluvial fans at the northern margin of the Hornelen basin, and found close correspondance with the sub-Devonian rocks that are presently exposed on Bremangerlandet. The Bremangerlandet rocks, however, is located on the Upper Plate, which constitutes the basin floor of the Hornelen deposits. Therefore, the source area of the Hornelen conglomerates was not the Bremangerlandet area *per se*, but similar rocks that were present in the source area to the north, east and south of the Devonian basin at the time of deposition, i.e. present in a geographical position that was part of the Lower Plate. These source rocks have later been removed by erosion, but remnants of the rocks are presently exposed on the basin floor at Bremangerlandet, showing that these rock types were also present at the Lower Plate surface during deposition of the Hornelen conglomerates. Now we leave the *conglomerates*, and return to the issue of source area for the *sandstones*. The provenance studies of Fonneland & Pedersen (2002b) implies that a “gneissic” source is required. Once again, the solution may be found at Bremangerlandet. Hartz et al. (1994) and Cuthbert (1991) have described a large gneiss area in these regions. Kildal (1970) correlated these Bremanger gneisses with the Western Gneiss Region. The gneisses are situated in central parts of Bremangerlandet, boardered to the south by the “outboard”/island arc complex represented by the Kalvåg Melange, the Bremanger Granodiorite, and various other rocks, and to the north by mylonitic gneisses and the Vetvika Shear Zone, which marks the northern boundary of the Upper Plate (**Fig. 2.3**). North of the Vetvika Shear Zone, we find mylonitic gneisses of the Nordfjord–Sogn Detchment Zone. According to Hartz et al. (1994), the Bremangerlandet gneisses have pre-Caledonian structures and consist of heterogeneous quartzofeldspatic rocks. Cuthbert (1991) reported that the Bremangerlandet gneisses contain no eclogites, and that all metabasic bodies are amphibolites with no signs of a previous eclogite mineralogy. The gneissic banding at Bremangerlandet is reported to be more coarsely differentiated than that found to the north of Nordfjord. Hartz et al. (1994) and Cuthbert (1991) reported that the Bremangerlandet gneisses were overlain by quartzites and schists, locally with a preserved unconformity. These metasediments were, based on lithological resemblance, correlated with similar rocks on the island of Atløy, and they had suffered only low-grade Caledonian meramorphism. The Bremangerlandet gneiss unit was tentatively enterpreted to occupy the same tectonostratigraphic level as the Dalsfjord Nappe and the Jotun Nappe. As we see from the referred publications, the presence of the Bremangerlandet gneiss proves that the Caledonian Lower-Plate *allochthons* that surrounded the Hornelen basin during deposition, did indeed contain gneisses. The Bremangerlandet gneiss itself, or more correctly, its possible counterpart on the Lower Plate/source area side, could have been the source, but it is equally possibly that other gneiss units in the Caledonian allochton was the source, units which have since been removed by erosion. Considering all these facts, it seems to be unlikely that the WGR was the source for the Hornelen sandstones. Rather, it appears likely that one or more allochthonous gneiss units in the Caledonian nappe stack served as the sediments source.

2.7.2 MAJOR FEATURES OF THE MASSIFS

2.7.2.1 SEDIMENTATION

The Devonian massifs generally contain marginally derived conglomerates which either fill the massifs entirely, or grade into axially transported sandstones in central or distal parts. The conglomerates form mass flow or streamflow-dominated alluvial fan and fan delta bodies, whereas the finer grained sediments commoly consist of braided alluvium, flood basin and lacustrine deposits (Steel et al. 1985).

2.7.2.2 CONTACTS

The contacts between the Devonian deposits and the substrates are of different types. In the west, the contacts are well-preserved primary unconformities. Along the southern and northern margins, the contacts are brittle faults, except in the Håsteinen Massif. Here the northern contact exhibits the primary unconformity a number of places despite minor faulting along the contact; and the southern margin is mostly a fault contact, but with the primary unconformity preserved in its western part (this work). Eastern margins for all the massifs are defined by fault contacts.

2.7.2.3 BEDDING GEOMETRY

The alluvial successions tend to form basin-wide and facies-independent cycles. These are particularly well displayed in the Hornelen Massif, where the megacycles (termed “cyclothemes” by Steel et al. 1985) are up to several hundred meters thick, and where the cycles may also contain lacustrine deposits. Generally, bedding has an

E-ward dip along the central axes of the deposits. The Solund Massif is an exception from this, with bedding dipping either towards NW or SE. The deposition took place in the eastern parts of the basins, and simultaneously the basin and the basin floor (Upper Plate) migrated westwards. Thus, the depocenters continuously migrated eastwards relative to the sediments that had alredy been deposited. Sequences have onlapping relationships towards the east (and southeast), resulting in large bedding-normal stratigraphic thicknesses, e.g. **25 km** in the Hornelen Massif (Steel et al. 1985).

2.7.2.4 BASIN FORMATION

The early models appearing in the literature suggested that the sedimentation in the basins involved propagating hinge faulting (Bryhni 1964b, 1978), and oblique-slip faulting with associated extensional faults across the basin floor (Steel et al. 1977; Steel & Gloppen 1980). Subsequently, various types of listric normal faulting have been applied (Bryhni 1982; Bjørlykke 1983; Hossack 1984), and these models have been further developed into the so-called "detachment model" (see Sect. 2.8).

2.7.2.5 DEFORMATION

The Devonian massifs have been deformed into open folds (e.g. Vogt 1928; Kolderup 1960; Roberts 1983; Seranne 1988; Chauvet & Seranne 1994; Osmundsen & Andersen 2001). The trend of the fold axes varies from NE-SW in the Solund Devonian Massif (Nilsen 1968; Sturt & Braathen 2001); WSW-ENE in the Kvamshesten Massif (Høisæther 1971; Bryhni & Skjerlie 1975; Osmundsen et. al. 1998; Braathen 1999); WNW-ESE in the Håsteinen Devonian Massif (Kolderup 1925; this work), and WSW-ENE in the Hornelen Massif (e.g. Bryhni 1964a; Krabbendam & Dewey 1998; Larsen 2002b). The fold axes always plunge in the eastward direction, except in Solund where the NE-SW trending fold axis is subhorizontal. In the overall picture, the folds are synclines, and the roughly E-dipping bedding along the axial trace give way to NE-ward dips along the southern margins and SW-ward dips along the northern margins. Again, the Solund Massif is an exception: it forms an anticline instead of a syncline, and with its NE-SW trending axial trace, it displays a southeastern limb dipping SE-wards and a northwestern limb dipping NW-wards. The mylonites below the Devonian massifs were traditionally interpreted as evidence of eastward thrusting of the massifs. This thrusting, and the folding mentioned above, have traditionally been ascribed to a late Caledonian orogenic event termed the Svalbardian Orogenic Phase (Vogt 1928; Roberts 1983), occurring in early Upper Devonian time (Vogt 1928). This deformation was termed the Solundian Orogenic Phase by Sturt (1983), and the model may be termed the “Solundian orogeny model”. From the mid 1980s, this model has, in the literature, been challenged by the "detachment model", which states that the contractional Caledonian Orogeny had ceased by the time the major volumes of Devonian sediments were being deposited. The model also states that the massifs were the result of large-scale detachments that formed due to crustal-scale extension of the orogen. This model opens for the possibility that folding of the Devonian deposits occurred during these overall extensional movements (Chauvet & Seranne 1994; Larsen 2002b). The new school of thought represented by the “detachment model” has been challenged by Sturt (1983), Torsvik et al. (1986, 1987, 1988) and Sturt & Braathen (1999, 2001) who argue for the existence of a Solundian/Svalbardian orogeny representing a late Caledonian phase that deformed the Devonian rocks (i.e. the “Solundian Orogeny model”). The "detachment model" is described in Sect. 2.8, and the "Solundian Orogeny model" in Sect. 2.9 below.

2.8 THE "DETACHMENT" MODEL

2.8.1 GENERAL FEATURES

2.8.1.1 ESTABLISHMENT OF THE MODEL

Bjørlykke (1983) was the first to publish the new idea that Devonian large-scale extensional detachments were of fundamental significance for the development of the geological architecture of western Norway, and that these detachments controlled the deposition of the Devonian basins there. Also Roberts (1983) considered these basins to have formed due to a phase of crustal extension, but viewed the basin-controlling faults merely as temporary structures that were later reversed to thick mylonitic top-to-the-E/SE thrust zones of late Caledonian age, and therefore did not appreciate the fundamental nature of the detachment zone.

Norton (1983: unpublished report) was inspired by the ideas of Hossack (see below), ideas which became published in Hossack (1984). Norton (op.cit.), who had studied the marginal contacts of the Devonian basins in western Norway, was the first to present field descriptions that demonstrated the extensional nature of both (i) the brittle contact faults at the base of the Devonian deposits, and (ii) the subjacent mylonites of the detachment zone.

Hossack (1984) was the first to outline the fundamental properties of this detachment system in western Norway; for example the large-scale displacements, and the geometries and distribution of the involved rock units. His work was based on close studies of the **1:250 000** geological map "Måløy" (Kildal 1970) that gives a fairly detailed overview of the area. The model that Hossack (1984) developed from these studies, the "detachment model", was thus based on the distribution and properties of mapped rock units.

In summary, the pioneer papers of Bjørlykke (1983), and in particular Hossack (1984), started a new school of thought concerning the geological development of western Norway, by introducing post-Caledonian Devonian crustal-scale extensional tectonics as the mechanism responsible for major elements of the rock architecture. The new concept implied that these extensional movements were considered to control features such as the basin formation, the sedimentation, and, according to Hossack (1984), also the tectonic deformation (and metamorphism) of the detachment system. Norton (1986, 1987) was the first to present *field data* documenting the detachment system, and the first to present models based on field data. Norton (1987) particularly documented the existence of the several km thick, crustal-scale mylonitic detachment zone that controlled the extensional movements in western Norway. Based on the geographical distribution of the zone, Norton (1987) named it the *Nordfjord–Sogn Detachment Zone*, and this has been the standard name for the zone in all subsequently published literature. The papers triggered a renewed interest in the geology of the area, and a number of authors have subsequently developed and refined the model, here for short called the "detachment model".

2.8.1.2 BRIEF OUTLINE OF THE MODEL

In plate tectonic terms, the detachment model can be briefly outlined as follows: Scandian continent–continent collision between Baltica and Laurentia in the Late Silurian−Early Devonian led to full-scale imbrication of the continental crust in western Norway. During this process, the Western Gneiss Region, that constituted the leading edge of Baltica was subducted below Laurentia and subjected to (ultra-)high-pressure (UHP) metamorphism, as recorded by the eclogites presently exposed in the region (Griffin & Brueckner 1980; Griffin et al. 1985; Cuthbert et al. 2000; Engvik & Andersen 2000; Carswell et al. 2003; Foreman 2005; Hacker 2007; Johnston 2007a). Allochthonous nappes of Baltic, outboard and Laurentian affinity were thrusted onto the Baltic craton. The cessation of contractional movements was accompanied by “collapse” of the isostatically unstable Caledonian nappe pile, which after some reverse movement along the original thrust décollement zone was cut by a major crustal-scale westerly-dipping detachment zone. This zone devided the region into a footwall plate in the east, and a hangingwall plate in the west. As the footwall plate rose due to buoyancy forces, thick sequences of Devonian coarse clastic continental Old Red Sandstone sediments were deposited on the down- and westward-moving hangingwall plate in the west (Norton 1986, 1987; Milnes et al.1988, 1997; Hacker 2007).

The large area presently situated between the coast of western Norway and the "Faltungsgraben" (Central Trough) can thus be divided into a *Lower Plate* (footwall) in the east, and an *Upper Plate* (hangingwall) in the west, separated by a low-angle westward-dipping *detachment zone* (**Fig. 2.7**) (Seranne & Seguret 1985; Michelsen 1986a; Norton 1986, 1987; Andersen et al. 1990; Hacker 2007), which has been named the “*Nordfjord–Sogn detachment zone*” (Norton 1986, 1987). The geological evidence for the model in each of these three regimes will be outlined below.

A review of HP/UHP terranes throughout the entire Scandinavia has been given by Brueckner & van Roermund (2004). The western Gneiss Region is **one of fifteen** areas around the world where UHP rocks have been found (Chopin 2003). The UHP rocks of the WGR, as well as the NSDZ, attract considerable international research, and are ranked among the best places to study such features. This has recently been expressed by everal authors: Terry & Heidelbach (2006, page 4): "*This window [northward from Nordfjord] has the best exposures of HP and UHP rocks in the world*”. Young et al. (2007, page 1): ”*The Nordfjord area of western Norway hosts one of the best known provinces of ultrahigh-pressure rocks in the world*”. Johnston et al. (2007a, page 10): “*The NSDZ is one of the largest extensional detachment zones on Earth*”.

2.8.2. LOWER PLATE (WESTERN GNEISS REGION, WGR)

2.8.2.1 INCREASING P/T CONDITIONS TOWARDS THE NW

In the present description of the Lower Plate, focus will be on the Western Gneiss Region. Along the eastern margin of this region, the gneisses disappear below the Jotun Nappe Complex and other Caledonian nappes of the "Faltungsgraben" (**Fig. 2.1, 2.2** and **2.7**). The NE-SW trending Lærdal–Gjende fault, which is positioned further east and which cut the Jotun Nappe Complex (Milnes & Koestler 1985; Milnes et al. 1988, 1997) (**Fig. 2.1, 2.2, 2.7**), may be defined as the proper eastern boundary of the Lower Plate, althought movements on this fault have been modest compared to movements on the Nordfjord–Sogn Detachment Zone (NSDZ). The Lærdal–Gjende Fault originated, together with the NSDZ in Middle Devonian times (Milnes et al. 1997), and palaeomagnetic dating (Andersen et al. 1999) showed that the fault was subjected to brittle reactivations in mid–late Permian and in late Jurassic–early Cretaceous times. The Lower Plate thus consists of both the WGR as well as a significant part of the Jotun Nappe Complex. In the following discussion, however, the term “Lower Plate” will refer to the Western Gneiss Region (WGR), since the rocks of interest are located and exposed here.

The easternmost areas of this Lower Plate contain migmatitic gneisses and granites of the Western Gneiss Region (Jostedalen Complex). These areas have not been affected by the Caledonian reworking (Milnes et al. 1988, 1997; Skår & Pedersen 2003), and do not seem to have experienced the Caledonian high pressure conditions (Cuthbert et al. 2000).

In contrast to these eastern parts, the areas further west- and northwestwards contain frequent occurrences of Caledonian eclogites and other high-grade rocks. Already by the early 1980s, radiometric dating had established that these rocks were of late Caledonian (Late Silurian–Early Devonian) origin (Griffin & Brueckner 1980, 1985; Griffin et al. 1985), with the protoliths having Gothian or Sveconorwegian ages. The rocks in this area were also found to become gradually more and more affected by Caledonian shearing and polyphase ductile deformation towards the west (Milnes et al. 1988, 1997; Cuthbert et al. 2000). Griffin et al. (1985) compiled eclogite thermometry data on a map and showed that NE-SW trending isotherms of **600**, **650** and **750°C** could be drawn parallel to the coast, with the highest temperature along the coast, the intermediate temperature further east, and the lowest temperature in the easternmost parts. On the map in Griffin et al. (1985), the **750°C** isotherm followed the line Bremanger–Ertvågøy (SE of Smøla); the **650°C** isotherm followed the line Inner Nordfjord–Surnadalen; and the **600°C** isotherm, which was divided into two segments, followed the line-segment Hyllestad–Førde in Sunnfjord, and, after an eastward step, continued as the line-segment Jostedalen–Dovre. The increase of temperatures towards the northwest demonstrated the presence of a regional isothermal gradient, and a corresponding *pressure* gradient also appeared to be present. This picture conforms with the model of deep subduction of the Western Gneiss Region towards the west and northwest. Later publications on thermobarometry have strengthened the model, with the presence of ultra-high pressure rocks in the NW (e.g. Hacker 2007; Johnston et al. 2007a).

2.8.2.2 P/T CONDITIONS FOR UHP AND HP ROCKS IN THE NW PART OF THE WGR

Smith (1984) was the first to find coesite associated with eclogites in the Nordfjord–Stadtlandet area, implying that ultrahigh-pressure (UHP) eclogites were found for the first time in the Western Gneiss Region – and in Norway. A more detailed review of the discovery was given in Smith (1988). From Stadtlandet, Wain (1997) found more localities of coesite, or polycrystalline quartz after coesite, increasing the number of UHP localities to **24** at the time. For some years to come, the UHP rocks were only found in the Stadtlandet region. Eventually, Dobrzhinetskya et al. (1995) reported microdiamonds from the Island of Fjørtoft north of Ålesund, **70 km** to the northeast of Stadtlandet, showing that the UHP rocks were not confined to Stadtlandet, but present also in a larger area. After these discoveries, an intensive search for UHP rocks has been carried out to find the extent of the UHP province, and numerous UHP localities have been described. For example, UHP rocks have subsequently been described from the group of islands called Nordøyane (**Fig. 2.1**) (of which the Fjørtoft island is a member) located north of Ålesund (Terry et al. 2000a, 2000b; Terry & Robinson 2003, 2004), and from the island Hareidlandet southwest of Ålesund (Carswell et al. 2003a). Walsh & Hacker (2004) found polycrystalline quartz after coesite as far east as Sunnylven/Hellesylt **(Fig. 2.1**) (at Geirangerfjorden), positioned **~85 km** east of the westernmost UHP locality Drage at Stadtlandet, thus expanding the UHP area considedrably eastwards. Also in the Stadlandet–Nordfjord area new discoveries were made, for example in the Selje–Maurstad region (Cuthbert et al, 2000), at Saltaneset (Carswell et al. 2003b), at Drage (Schärer & Labrousse 2003), and at Verpeneset (Root et al. 2004). From the archepelago of Sørøyane, Root et al. (2005) described new localities of UHP rocks, and reported that the total number of such localities in the WGR had by then reached **30**. These authors also suggested that the UHP rocks could now be assigned to **3** separate geographical areas which they termed **i)** the *Nordøyane UHP domain*, **ii)** the *Sørøyane UHP domain* and **iii)** the *Nordfjord–Stadtlandet UHP domain* (**Fig. 2.1**) In eastern Nordfjord, Young et al. (2007) found a new UHP eclogite at Heggjadalen north of lake Hornindalsvatn, a locality situated **65 km** east of Drage at Stadtlandet. The Heggjadalen locality implies that the Nordfjord–Stadtlandet UHP domain can be extended considerably eastwards, to regions southeast of Sunnylven, (the latter belonging to the Sørøyane UHP domain, Hacker 2007).

The distance from the Hornelen sediments to the UHP rocks is of intrest. The Heggjadalen locality is located **35 km** to the ENE of the northeastern tip of the Devonian massif. However, the UHP rock locality positioned *closest* to the Devonian sediments is found at Verpeneset (Root et al. 2004) in outer Nordfjord at a locality situated only **5 km** north of the Hornelen Devonian sediments (the very Hornelen mountain) of Bremangerlandet. Across this distance of only **5 km,** a transition can be followed from Devonian deposits that experienced a maximum metamorphism of lowest greenschist facies at **10 km** burial, to UHP rocks that experienced

UHP metamorphism at **100–140 km** depth – demonstrating that a crustal section of **~100 km** or more is missing between the two units. The two units are sepatated by the Nordfjord–Sogn Detachment Zone (NSDZ) that in the area trends E-W, and that appeares to have played a role in the juxtaposition of the two units. The adjacent location of the two units is an intriguing documentation of excision of extremely thick portions of crustal section. The NSDZ are further described below.

Estimates of maximum pressure and temperature of UHP and HP rocks have been reported from several areas. From the Island of Fjørtoft (Nordøyane domain), for example, Terry et al. (2000a) performed analyses on UHP rocks that yielded **T = 820°C** and **P = 34–39 Kbar (= 3.4–3.9 Gpa)**, indicating a maximum crustal depth of at least **125 km**, possibly as much as **140 km** (Young et al. 2007). When generally considering the UHP areas, i.e. also including the Stadtlandet–Nordfjord domain, the pressure and temperature for *all* UHP areas fall in the range **T = 600–820°C and P = 28–39 Kbar** (Wain 1997; Cuthbert et al. 2000; Terry et al. 2000a). The UHP areas are surrounded by HP areas. Pressure estimates of such HP rocks from the areas immediately to the south of the mixed HP/UHP zone (see below) at the Stadtlandet–Nordfjord domain have yielded values in the range of **20–25 Kbar (= 2.0–2.5 GPa)** (Cuthbert et al. 2000; Krogh Ravna &Terry 2004; Labrousse et al. 2004), indicating a maximum crustal depths of **c. 75 km**. In a study from inner Nordfjord, Young et al (2007) analysed samples taken from scattered eclogite pods mainly at Sandane and Stryn (**Fig. 2.1**), and obtained **P = 2.2–2.5 GPa, T = 593–617°C**, which is also in the same range.

The Stadtlandet and Nordøyane domains have received paricular attention in the discussion on the mechanism for juxtaposition of the UHP rocks with the HP rocks. At Stadtlandet, Krabbendam & Wain (1997) mapped the extent of the UHP rocks, and found that a <u>boundary zone</u> separated the westerly located UHP rocks from easterly located HP rocks. Wain (1997) advocated that a pressure difference of **>4 Kbars** existed between the UHP and HP rocks, and speculated that the UHP rocks were separated from the HP rocks by a tectonic contact. Based on work at Nordøyane north of Ålesund, Terry et al. (2000b) suggested that the UHP had been thrusted SE-wards on top of the HP rocks during the time interval **407– 401 Ma**, and that UHP and HP rocks together had been thrusted further until **395 Ma**. From Stadtlandet, Cuthbert et al. (2000) confirmed the presence of a transitional zone between UHP and HP rocks, but the model of thrusting was generally questioned by Cuthbert et al (2000), arguing that it still remained to be found convincing tectonic/structural features indicating tectonic movements in the **10 km** wide <u>mixed zone</u> (= boundary zone). Terry & Robinson (2004) presented more work from Nordøyane, and strengthened their argumentation for eclogite facies thrusting. Still from Nordøyane, notably from the island Haramsøya, Terry & Heidelback (2006) presented details of shear zones that were interpreted to have developed during HP to UHP conditions. From Sørøyane, which is located **30 km** southwest of Nordøyane, Root et al. (2005) advocated that the UHP rocks are positioned <u>*underneath*</u> the HP rocks, and that the UHP terrane is exposed in the centre of regional E-W trending anticlines. The UHP rocks were interpreted to constitute one single large unit with considerable crustal thickness. Root et al. (2005) further argued that since the thrusting advocated by Terry & Robinson (2004) can, in the field, only be observed locally, whereas the contact elsewhere is invisible and identifiable only by petrology and chronology, the upper contact against the HP rocks must have formed “prior to the final granulite facies metamorphism, which must have annealed any shear zone along the contact”. Hacker (2007), in a review of published radiometric and P/T data, supported the view that UHP rocks are *underlying* the HP rocks. This was also advocated by Johnston et al. (2007a) in their review of the region between Solund Devonian Massif and Sørøyane, a N-S distance of **160 km**.

From field work to the southeast in the WGR, i.e. in the area just north and east of the Hornelen Devonian Massif, Young et al. (2007) showed that amphibolite facies rocks found in the SE were, within 20 km in the NE-ward direction, changing gradually via HP rocks into the UHP rocks in the NW (northern part of Nordfjord). This shows that these parts of the WGR contained, at the time when WGR was a subducted slab, UHP rocks that updip were attached to lower pressure rocks, i.e that the (U)HP and lower-P rocks did not become juxtaposed by tectonic movements. Hense, the authors concluded that these parts of the WGR were exhumed as one large slab unit. Hacker (2007) and Walsh et al (2007) suggested that *all* the UHP areas of the WGR were exhumed as part of one, single, relatively coherent slab.

2.8.2.3 P/T CONDITIONS FOR HP ROCKS IN SUNNFJORD AND SOGN

In the Sunnfjord area, detailed studies have recently been conducted in eclogitic gneisses around Dalsfjorden south of the Kvamshesten Devonian Massif, notably at the headland of Vårdalsneset (Andersen et al. 1994; Engvik & Andersen 2000) (**Fig. 2.3**), at the island Bårdsholmen (Engvik et al. 2000), and in the Holt–Tyssedalsvatnet area on the mainland south of Bårdsholmen (Engvik et al. 2001) (**Fig. 2.3**). P-T analyses from Vårdalsneset (Engvik & Andersen 2000) yielded **T = 690 +/- 20°C, P = 16 +/- 2 Kbars**. From samples collected at Sellevoll south of Dalsfjorden (**Fig. 2.3**), Cutbert et al. (2000) reported **T = 580°C, P = 21 Kbars**. The Drøsdal eclogite (**Fig. 2.3**) south of Dalsfjorden was investigated by Forman et al. (2005), who obtained **T = 720–830°C, P = 19–21 Kbars**. These results comply with P-T estimates in earlier works from Sunnfjord, where pressures were reported to be up to **18 Kbar** and temperatures were in the range **500–700°C**, (Krogh 1980), although the maximum temperature of the Drøsdal eclogite is somewhat higher than the latter interval.

Further south, i.e. in outer Sogn, P-T studies were carried out by Chauvet et al. (1992) and Hacker et al. (2003), both focusing on the Hyllestad–Lavik area east of the Solund Devonian Massif **(Fig. 2.3**). The geology of the Hyllestad area was detailed by Tillung (1999). Chauvet et al. (1992) studied schists and eclogites, and reported pressures of **11–15 Kbar**, and temperatures of **500–600°C**. Hacker et al. (2003) reported that the Hyllestad Complex gave a temperature range of **575–670°C** and a pressure of **1.5 GPa (=15 Kbar)**, whereas the eclogites that were situated slightly further east yielded a temperature of **c. 700°C** and a pressure of **2.3 GPa (=23 Kbar)**. In summary, the *temperature* of the HP rocks of outer Sogn (**500–600°C, 575–670°C, ~700°C**) is thus consistent with that in HP rocks of Sunnfjord (**500–700°C**) further north. Also the pressures seem to be overlapping, with **11–23 Kbar** for Sogn and **16–21 Kbar** for Sunnfjord, although the lower limit of **11 Kbar** pressure obtained by Chauvet et al. (1992) for outer Sogn (**11–15 Kbar**) is lower than the **15 kbar** that Hacker et al. (2003) obtained for the same area.

2.8.2.4 RADIOMETRIC DATING OF PEAK METAMORPHISM AND SUBSEQUENT UPLIFT

The age of the eclogite peak metamorphism of the WGR has traditionally been estimated to **ca. 425 Ma** (**425 +/- 12 Ma**), an age which was based on the Sm-Nd analyses of garnet-omphasite from **5 eclogites**, reported by Griffin & Brueckner (1980, 1985). At the time of publication, this age was important (Carswell et al. 2003a), because it corresponded to Caledonian mineral ages in the surrounding gneisses, and thus showed that the eclogites were Caledonian and not Precambrian as previously argued in the literature (e.g. Krogh 1977), and because it showed that the eclogites had formed ***in situ*** in the crustal rocks, and had not originated in the mantle. However, according to Carswell et al. (2003b), who re-examined the dating, error bars on this ***mean*** age of **ca. 425 Ma** was **larger than 80 Ma** rather than the firstly estimated **+/-12 Ma** when error brackets on each of the involved age determinations were considered.

In subsequent radiometric studies, the dating of formation of the UHP eclogites have been of particular interest, and the following ages have been obtained in the three UHP domains:

From the *Nordøyane domain*, early studies of the UHP event at the island Flemsøya yielded **410 +/- 16 Ma** (three point Sm-Nd isochron, Mørk & Mearns 1986)**.** Newer datings of the UHP event gave **415 +/- 7 Ma** (Pb/U monazite, Terry et al. 2000b) from the island Fjørtoft (located **22 km** north of Ålesund). Krogh et al. (2004) reported the three ages **415 +/- 1 Ma** (U/Pb zircon) from the island Averøya, **410 +/- 1 Ma** and **408 +/- 1 Ma** from Flemsøya, and **411.5 +/- 1.2 Ma** from the island Lepsøya.

From the *Sørøyane domain*, the age **401.6 +/- 1.6 Ma** (U/Pb, zircon) was reported by Carswell et al. (2003a) from the Ulsteinvik–Dimnøy area at the island Hareidlandet (**17 km** southwest of Ålesund. The identical age **401.6 +/- 1.6 Ma** (U/Pb, zircon) was subsequently obtained by Tucker et al. (2004) for the eclogite at Ulsteinvik on the island Hareidlandet.

From the *Nordfjord–Stadtlandet domain*, an early study by Mearns (1986) gave the age **408 +/- 0.8 Ma** (three point Sm-Nd isochron) from Almklovdalen. Carswell et al (2003b) reported the age **408.3 +/- 6.7 Ma** (Sm-Nd, multipoint garnet, clinopyroxene, wholerock) from Saltaneset. From Flatraket, Root et al. (2004) reported eclogite zircon recrystallisation (U/Pb) in the interval **405–400 Ma**, and at Verpeneset, **5 km** north of the Hornelen Devonian sediments on Bremangerlandet, they reported the U/Pd-zircon age **404 +/- 21 Ma**. Young et al (2007) obtained the age **405 +/- 2 Ma** (U/Pb, zircon) from Bjørkedalen north of lake Hornindalsvatn.

As noted by Young et a.l (2007), the ages of the Nordøyane domain mainly fall in the interval **410–415 Ma**, whereas the ages of the Sørøyane and Nordfjord–Stadtlandet domains appear to form a group concentrated in the interval **405–400 Ma**. However, as the error bars on some of the ages of the two groups are overlapping, it is at present uncertain whether the ages do actually form two groups (this work).

The **405–400 Ma** interval of the Nordfjord–Stadtlandet and Sørøyane UHP eclogites partly overlaps the **402–394 Ma** age interval reported by Fossen & Dunlap (1998) for the extensional, westward movement of the Caledonian orogenic wedge on its subjacent décollement zone. The two processes; i.e. development of peak eclogite metamorphism, and incipient extensional movements of the orogenic wedge, may have occurred at about the same time, and are thus not in conflict (this work). When considering ages of peak eclogite metamorphism, it should be noted that this metamorphism was not coeval everywhere in the WGR. At Drage on the peninsula of Stadtlandet, for example, Schärer & Labrousse (2003) reported formation of eclogites as late as **389 +/- 2 Ma** (U/Pb, rutil-omphacite). At the time of publication, this was the youngest age obtained for formation of eclogite in Western Norway. The age was interpreted to represent freezing of the eclogite mineral paragenesis at maximum *temperature*, estimated to **770°C** at a not maximum pressure of **1.8 GPa**. The actual maximum *pressure* experienced by the rock, estimated to **2.8 GPa**, occurred prior to this, but the age of this maximum pressure could not be found since radiomatric dating of coesite and diamonds is impossible (Schärer & Labrousse 2003). Young eclogite ages were also obtained by Kylander-Clark et al. (2007) who used the Sm-Nd method, and also – for the first time in the WGR – the Lutenium–Hafnium (Lu-Hf) method. The ages **380 +/- 14 Ma** (Lu-Hf) and **388 +/- 10 Ma** (Sm-Nd) were obtained from the Otrøy island in the Nordøyane UHP domain, and the ages **369 +/- 11 Ma** (Lu-Hf) and **484 +/- 11 Ma** were obtained from the Remøya island in the Sørøyane UHP domain. (Other, more “normal” eclogite ages were also reported). The age **369 +/- 11 Ma** (Lu-Hf) is the hitherto youngest age recorded for formation of eclogite in Western Norway. It was noted that the ages overlap $^{40}Ar/^{39}Ar$ ages from the same or near by localities. The young eclogite ages were interpreted to reflect cooling at eclogite facies conditions, and *not* prograde garnet growth, nor peak metamorphic conditions. Despite these young ages, the general peak meatmorphism for the UHP eclogites in WGR was still considered to be around **~405–400 Ma**, and the prograde phase leading up to this peak to have started at **~425–420 Ma**.

The peak metamorphism was succeeded by a post-eclogitic metamorphic retrogression, caused by rapid isothermal uplift of the region, that produced an amphibolite-facies metamorphism in large parts of the Western Gneiss Region (e.g. Griffin et al. 1985; Cotkin et al. 1988; Dunn & Medaries 1989; Chauvet et al. 1992; Dewey et al 1993; Milnes et al 1997; Engvik & Andersen 2000; Terry et al. 2000; Root et al. 2005; Young et al. 2007). The uplift presumably reflected the extension-related buoyancy of the area (Andersen et al. 1991; Austrheim 1991; Andersen et al. 1994; Milnes et al. 1997; Hacker 2007). From the western part of the “Sognefjorden log”, Milnes et al. (1997) demonstrated that the uplift of the eclogites had occurred under pure shear conditions. These obervations was supported by Marques et al. (2007), who in a study of the structural development of the NSDZ found pure shear at the base of the zone.

In the WGR north of the Hornelen Devonian deposits, Lux (1985) was the first to apply K/Ar and $^{40}Ar/^{39}Ar$ geochronology to record the late stages of eclogite uplift. This study dealt with the Stadtlandet area, and the $^{40}Ar/^{39}Ar$ dating of hornblende, sampled from Nordpollen 10 km east of Måløy town showed that the rocks cooled through the hornblende blocking temperature (T_B) of **500–550°C** at **410 +/- 1 Ma**, whereas $^{40}Ar/^{39}Ar$ dating of biotite from Hoddevik at the western end of the Stadtlandet peninsula indicated cooling through its T_B of **300–350 °C** at **375 +/- 6 Ma**. Hense, the study documented for the first time, the uplift from eclogite conditions to amphibolite facies and further to lower greenshist facies conditions. It should be noted, however, that the hornblende age of **410 +/- 1 Ma** may seem to be too old when compared to recent dating of the Nordfjord–Stadtlandet eclogite formation (see above) that seems to have peaked in the interval **405–400 Ma** (Young et al. 2007). Although the error bars on several of the eclogoite datings are wide, the ages seem to indiate that the amphibolite facies metamorphism must have taken place after **~400 Ma**. Dating of the amphibolite facies from other parts of the WGR seems to support such a later age for this metamorphism. For example, as also noted by Kylander-Clark et al. (2007), radiometric ages of the amphibolite facies have beeen obtained from Sørøyane–Averøya (zircon from pegmatites); from the Fjørtoft island at Nordøyane (monazite, Terry et al. 2000b); and from regional sampling of titanite (Tucker et al. 1990; 2004), suggesting that the amphibolite facies metamorphism in the WGR occurred in the interval **395–390 Ma**. However, as also noted by Walsh et al (2007), the age of the amphibolirt facies metamorphism in the WGR is somewhat poorly constrained, suggesting that caution should be executed when dealing with the subject. A study by Andersen (1998), from the outer part of Nordfjord, reported numerous $^{40}Ar/^{39}Ar$ muscovite plateau ages that fell in the range of **390–385 Ma** (**T_B = 350–400°C**). From Sørøyane, Root et al. (2005) reported partly overlapping $^{40}Ar/^{39}Ar$ muscovite ages in the interval **387–379 Ma** (recalculated in Walsh et al. 2007). The ages of Andersen (1998) appear to fit well with the other mentioned exhumation timing data from the Stadtlandet–Nordfjord area: Stadtlandet had the deepest burial at eclogite facies level at **405–400 Ma**; Nordfjord recided at **350–400°C** at **390–385 Ma**, and Stadtlandet reached the **300°C** level at **370 Ma**.

However, when discussing the uplift history of the Stadtlandet area, the above mentioned young eclogite formation at **389 +/-2** Ma (Schärer & Labrousse 2003) at Drage, Stadtlandet (located **12 km** SE of the western tip of the peninsula), must be considered, since the age indicates that Stadtlandet remained in eclogite conditions considerably longer than what is implicit in the above indicated exhumation timing. In addition, more young eclogites have recently been discovered also further north: Kylander-Clark et al. (2007) obtained the ages **380 +/- 14 Ma** (Lu-Hf) and **388 +/- 10 Ma** (Sm-Nd) from the Otrøy island in the Nordøyane UHP domain, and the ages **369 +/- 11 Ma** (Lu-Hf) and **484 +/- 11 Ma** from the Remøya island in the Sørøyane UHP domain. The age **369 +/- 11 Ma** is so far the youngest eclogite age recorded in the WGR.

One very interesting aspect of the young eclogites is that they apparently formed subsequently to the onset of the top-to-the-west Mode-I movement of the Caledonian orogenic wedge along the basal decollement zone, dated at **402–394 Ma.** (Fossen & Dunlap 1998). If such a westward movement of the entire nappe pile is termed post-Scandian, which is reasonable when assuming that the movement reflects plate divergence between Laurentia and Baltica, this late eclogite formation can also be termed post-Scandian. The implications imposed by the young eclogites for the exhumation history of these parts of the WGR is not clear. The lateral distance between the sample locality of Drage, Stadtlandet, that gave eclogites at **389 +/-2 Ma,** and the locality of Hoddevik that yielded ~**300°C** at about **370 Ma** (Lux 1985) is only **5 km**. Axcess argon in the samples dated by Lux (1985) could be one possibility, although Lux (1985) argue that the ages are not affected by excess argon. Nevertheless, as noted by Schärer & Labrousse (2003), the dated eclogite formation at **389 +/-2 Ma** in the Stadtlandet area was contemporaneous with amphibolite facies conditions in other segments of the Western Gneiss Region. This young eclogite age even overlaps with the age of the lower greenschist facies ($^{40}Ar/^{39}Ar$ muscovite) of Nordfjord. One important factor in the timing of the uplif, is the diachroneity in the process: during the exhumation of the subducted, westward-dipping WGR slab, the eastern parts would tend to reach the different isotherms earlier than the western parts. More research is needed to resolve the overlapping ages.

The similar type of retrogression to amphibolite facies has also been reported from the socalled *Vestranden area*, the continuation of the Western Gneiss Region to the northwest of Trondheim (Dallmeyer et al. 1992).

2.8.3 DETACHMENT ZONE (NORDJORD–SOGN DETACHMENT ZONE, NSDZ)

2.8.3.1 GEOGRAPHICAL EXTENT AND NAMING

At the western edge of the Lower Plate (**Fig. 2.7**), the sheared character of the rocks is dramatically enhanced, and a several km thick mylonitic to ultramylonitic belt is present along the coast. In map view, the belt is sinuous-shaped with alternating frontal and lateral ramps **(Fig 2.3** and **2.8)**. The mylonite belt is interpreted as part of the "Nordfjord–Sogn Detachment Zone" (see below) that was formed between the Lower Plate and the Upper Plate. The zone was first termed "the Måløy Fault" by Hossack (1984), who assumed that the structure was a single fault. Norton (1986) changed the name of the fault to the “Nordfjord–Sogn Detachment” and on Fig. 7 of Norton (1986) it was indicated that this name was meant to designate the several kilometre thick mylonitic *shear zone*, and not the brittle *fault* on top of the shear zone (compare with Fig. 5 of Norton 1987). Norton (1987) explicitly stated that the name "*Nordfjord–Sogn Detachment*”, should designate the kilometre-thick mylonitic detachment zone.

In the present thesis, the belt of mylonites are named the “Nordfjord–Sogn Detachment *Zone*” (NSDZ). The word “Zone” is added to avoid mix-up with the brittle *fault,* which is often named the “Nordfjord–Sogn Detachment” in the literature, but which is a structure that formed later.

In the study-area of the present work, both the Eikefjord Group and the Lykkjebø Group are extensively mylonitized (Wilks & Curthbert 1994; Johnston et al. 2007b; own observations), and these groups have also previously been interpreted as part of the detachment zone (Andersen & Jamtveit 1990; Wilks & Cuthbert 1994; Vetti 1997). The whole area between the Haukå Fault (**Fig. 2.6**) at the southern margin of the Hornelen Devonian Massif and the Sunnar Fault to the north of the Håsteinen Devonian Massif, is interpreted to have been mylonitised within the detachment zone (Andersen & Jamtveit 1990; Wilks & Cuthbert 1994; Vetti 1997) **(Fig. 2.3)**.

2.8.3.2 STRUCTURAL AND METAMORPHIC DEVELOPMENT

The crustal-scale west-dipping low-angle shear zone is supposed to have developed to accommodate the isostatic adjustments that became necessary when the Caledonian contraction ceased (see below). At the time when the Nordfjord–Sogn Detachment Zone (NSDZ) formed, the zone cut through the Caledonian nappes and reached the subaerial surface somewhere between the present-day exposures of the detachment zone and the Jotun Nappe Complex to the east. The Caledonian allochthons that lay on top of the WGR at this time, east of the present surface trace of the NSDZ, have later been eroded away. When the NSDZ formed, it reached a depth of at least **~20 km** below the surface, as evidenced by the amphibolite facies shear fabric described by Wilks & Cuthbert (1994) from the NSDZ mylonites between the Hornelen and Håsteinen Devonian Massifs, a fabric that was retrograded to greenshist facies during later stages of the shear movements. (The depth of **20 km**, and a thermal gradient of for example **30°C/km**, would correspond to a temperature of **600°C**, i.e. lower amphibolite facies). Recently, Johnston et al. (2007b) performed thermobarometric investigations in the shear zone, suggesting that the shear initiated even deeper, at **30–40 km** depth. Near the subaerial surface, the NSDZ had the character of a brittle fault, gradually changing to a thick mylonitic detachment zone at depth. To the west of the surface trace of the zone, the extension led to relative downward movement of the Upper Plate and formation of Devonian basins on its surface (**Fig. 2.7**).

Several authors have studied shear sense indicators in various segments of the detachment zone (or in the adjacent gneisses of the Lower Plate), and have reported consistent top-to-the-west sense of shear (**Fig. 2.8**) (Norton 1986,1987; Seranne & Seguret 1987; Chauvet & Seranne 1989; Andersen & Jamtveit 1990; Swensson & Andersen 1991; Wilks & Cuthbert 1994; Bøe 1997; Krabbendam & Dewey 1998; Tillung 1999; Young et al. 2007b; Johnston et al. 2007b). Along westward traverses across N–S trending segments of the mylonite belt, the metamorphic grade of the rocks has been found to show gradual retrogression from the eclogite-granulite facies in the east, via amphibolite facies to greenschist facies to the west (Michelsen 1986; Seranne & Seguret 1987; Chauvet & Brunel 1988; Chauvet & Seranne 1989; Chauvet et al. 1992; Tillung 1999; Hacker et al. 2003). Andersen & Jamtveit (1990) divided their Kvamshesten and Standalen Detachment segments into three zones corresponding to deep-, middle- and upper-crustal metamorphic conditions. In a detailed study of the Standalen Detachment, Krabbendam & Dewey (1998) did not use the zonal subdivision suggested by Andersen & Jamtveit (1990). Later workers have also avoided this subdivision, appaently because it is difficult to apply in the very heterogeneously developed Nordfjord–Sogn Detachment Zone.

Johnston et al (2007b) studied the NSDZ segment between Eikefjord and Hyen/Sandane, and reported on the structural, thermobarometric and geochronologic development. Their segment, which they called the "Hornelen segment", appeared to have developed in the time interval **410–400 Ma**. As indicated above, they suggested that the movements initiated at **30–40 km** depth, and that most of the movements also occurred during these amphibolite facies conditions. The shear was found to have been distributed across the entire thickness of the zone. The NSDZ was suggested to be responsible for the exhumation of the UHP rocks and the rest of the WGR slab, but only from the base of the crust, not from the mantle. A review of Johnston et al. (2007b) is given in Ch. 5.1.

2.8.4 UPPER PLATE

The areas west of the Nordfjord–Sogn Detachment Zone constitute the Upper Plate (**Fig. 2.7**), which again can be divided into two main units; the Caledonian nappe pile (Sect. 2.8.4.1) and the unconformably overlying Devonian sediments (Sect. 2.8.4.2).

2.8.4.1 CALEDONIAN NAPPE PILE

The Caledonian nappe pile may in its turn be devided into two units ; the Lower Palaeozoic rocks (se Sect. 2.3, 2.5 and 2.6), and the subjacent Dalsfjord Nappe, etc., consisting of Proterozoic "Jotun-Kindred" igneous and gneissic rocks (dated to **1634 +/- 3 Ma** by Corfu & Andersen 2002) with unconformably overlying Late Proterozoic continental margin sediments. **(Fig. 2.3)**. The Upper Plate rocks have suffered only low-grade greenschist facies metamorphism during the Scandian orogenic cycle (Brekke & Solberg 1987; Andersen & Jamtvedt 1990; Andersen et al. 1998; Eide et al. 1999). The spectacular juxtaposition of eclogites of lower to double crustal origin (**60-80 km**) in the Lower Plate against greenschist facies rocks of shallow crustal origin (**10-20 km**) in the Upper Plate, on either side of the detachment zone (**Fig. 2.7**), is, as illustrated above, elegantly explained by the "detachment" model.

2.8.4.2 DEVONIAN ROCKS

The Devonian rocks are resting unconformably on top of the Caledonian Nappe pile of the Upper Plate.

Basin formation

The basin formation has been explained as a result of the relative down- and westward movement of the Upper Plate during the uplift of the Lower Plate (Hossack 1984; Michelsen 1986; Seranne and Seguret 1987; Andersen & Jamtveit 1990**;** Milnes et al. 1997; Osmundsen et al. 2001; Johnston et al. 2007a). The uplift process led to continuous development of relief in the east, and as a result the ORS was deposited in enormous thicknesses in basins on the Upper Plate, constantly receiving sediments from the uprising source areas of the Lower Plate (Norton 1987). These source areas, which contained Caledonian nappes, have later been removed by erosion, as evident from the present-day exposed basement gneisses of the Western Gneiss Region. As mentioned above, rocks similar to those that were present in the source areas on the Lower Plate at the time of basin formation, now constitute the Upper Plate “basin floors” below the Devonian massifs. If we use the Hornelen massif as an example, the Upper Plate Bremangerlandet area now display the rock types that at the time of basin formation continued into the Hornelen source area on the Lower Plate side of the basin margin faults. Since the Bremangerlandet rocks represented the basin floor, they were soon burried by Devonian sediments when the basin formed, and thus the Bremangerlandet rocks were *themselves* never a source for the Hornelen sediments.

Eastward dip of strata

The eastward dip of the strata in the Devonian massifs has been explained by bedding-rotation resulting from the listric character of the eastern end of the detachment (e.g. Seranne & Seguret 1987; Osmundsen et al. 2000). From the southern margin of the Hornelen basin, Steel (1988) has described a diachronous belt of offset-stacked marginal fan bodies which confirm a westward movement of the basin during deposition, in accordance with the "detachment" model (Steel 1988).

Metamorphism

The anchizone to lowermost greenschist facies metamorphism in the massifs (Seranne & Seguret 1987; Torsvik et al. 1986; Sturt & Braathen 2001), with temperatures up to **300 +/- 50°C**, has previously been related to a sedimentary burrial of **5–13 km** (e.g. Seranne et al. 1989; Svendsen et al. 2001), depending on the assumed heat flow in the area. From studies of fluid inclusions in veins, Svendsen et al (2001) found that the present Solund basin had experienced a temperature of **305–330°C** and a burrial depth of **13 km**, and that the Hornelen and Kvamshesten basins had experienced a temperature of **250 +/- 20°C**. No data existed from the Håsteinen massif.

Folding

The folding of the west Norwegian Devonian massifs into E-W trending folds, was a result of contraction that in the literature has been interpreted in two different ways: the folding occurred either (i) *during* the extensional movements (e.g. Hossack 1984; Norton 1987; Chauvet & Seranne 1989, 1994), or (ii) *subsequent to* these movements (Braathen 1999; Braathen & Sturt, 2001).

Of the two alternatives, a recent study by Larsen (2002b) found support for alternative (i): Larsen (2002b) carried out detailed studies at the <u>northern margin</u> of the Hornelen basin, and found that the folding in this area (Ålfoten monocline) was entirely a result of intra-basinal processes, and not due to external regional contraction. The folding was interpreted as a result of <u>transpression</u> towards the basin margin *during* the extensional movements of the basin during its formation. These deformational processes *within* the basin were found to be decoupled from deformation of the gneisses *north* of the basin, where the folding was found to result from constrictional N-S shortening and E-W extension. Also the margin-parallel folding along the Hornelen <u>southern margin</u> (Grøndalen syncline) was suggested to be a result of the westward movement of the basin, but here the basin was believed to be in a <u>transtensional</u> setting. It was speculated that the folding along the southern margin accommodated some underlying geometry of a sinistral/normal boundary fault.

Another subject debated in the literature has been the question of wether the folding occurred whilst the sediments were in an unconsolidated (soft) state (Chauvet & Seranne 1994), or whether it happened after the sediments had become well lithified. Larsen (2002b) analysed veins that formed during the folding of Hornelen, and found that the veins were cutting clasts and matrix alike, showing that the veins formed in well lithified sediments, and implying that the *folding* had affected well lithified sediments also.

2.8.5 FOLDING OF THE NORDFJORD–SOGN DISTRICT

In western Norway, folding has also affected the units *below* the Upper Plate, i.e. the Nordfjord–Sogn Detachment Zone and the Lower Plate. N-S contraction has produced the E-W trending folds, and as shown below, various models have been proposed to explain the folding.

2.8.5.1 FOLD MODELS RELATED TO TECTONIC EVENTS

Briefly reviewed, the folding of the Nordfjord–Sogn district has been suggested to be a result of:

Model no. 1: Avalonian contractional forces (External forces):

Seranne et al. (1991) and Chauvet & Seranne (1994) argued for this model, where the docking of the northward-moving Avalon block (= British areas south of the Iapetus suture; + Germany; etc.), against Baltica in the north, were believed to have contributed to the formation of a large-scale sinistral shear system that gave transcurrent movements along the large British faults (e.g. Iapetus Suture zone, Highland Boundary Fault, etc) and the Møre–Trøndelag Fault Zone. This sinistral shear was believed to have caused the N-S folding in western Norway. However, as documented in the literature (e.g. Torsvik et al. 1996; Torsvik 1998; Cocks & Fortey 1998; Cocks & Torsvik 2002), Avalonia amalgamated with Baltica already by the **end of Ordovician or earliest Silurian**, **(i.e. around 443 Ma,** time-scale of Gradstein & Ogg 1996; Gradstein et al 2004), implying that the Avalonia collision could not have folded the Devonian basins which were deposited as late as the **Middle Devonian (391–370 Ma**, Gradstein & Ogg 1996; **398–385 Ma**, Gradstein et al. 2004), i.e. at least **40 Ma** after the collision. Consequently, this model has been abandoned in the literature.

Model no. 2: Hercynian (Variscan) contractional forces (External forces):

This model suggests that the folding was caused by contractional forces imposed by the Hercynian (Variscan) orogenic phase, which affected the European continent (Germany, Belgium, France) and the southernmost parts of Britain in the time period from the **Middle Devonian through Carboniferous**. The Hercynian/Variscan orogeny was a result of the collision of the continent Laurentia–Baltica–Avalonia with the continent Gondvana (Africa), a collision which included the amalgamation of the microcontinents of the Armorica–Ibrian Massif and the Bohemian Massif, that were located in the ocean between the two large continents (Torsvik et al. 1996). Sturt (1983) proposed that the Hercynian/Variscan contraction was responsible for the Solundian folding of western Norway.

This model is highly problematic, since the E-W folds, etc., of western Norway have an areal distribution confined to the *northern* half of the district. If this folding was a result of Herzynian contraction, the deformation intensity should generally increase southwards in western Norway, i.e. in the direction towards the central parts of the Herzynian / Variscan orogenic belt. This is not the case; rather the folding appears to be absent in the southern part of western Norway. Moreover, when taking into consideration the fact that Brittain was only affected in the souternmost part, it appears unlikely that the contraction should affect the whole of western Norway, reaching as far north as to the Trondheimsleden. This model of Hercynian/Variscan contraction has not gained much support in the literature.

<u>Model no. 3</u>: Svalbardian / Solundian orogenic phase (External forces):

From studies early in the 20th century, the Svalbardian Orogeny was established by Vogt (1928), to explain the deformation of the Devonian rocks of Spitsbergen, Bear Island (Bjørnøya), Western Norway, and also Scotland. The orogeny has been the traditional explanation for the N-S contraction in western Norway, but the model has also been discussed in recent works (Sturt & Braathen 2001). According to Vogt (op.cit.), the orogeny took place in the **lower part of Upper Devonian**. The <u>maximum age</u> limit for this dating was given by the fact that the folded Devonian rocks of Spitzbergen contained *fossils* that were assigned to the transition between the Middle and Upper Devonian, suggesting that the deformation occurred at this time or later. The minimum age was given by the fact that the folded layers were unconformably overlain by an undeformed, subhorizontal sedimentary sequence, that on Spitsbergen was of Lower Carboniferous age. A better minimum age, however, was obtained by correlation with the Bear Island (Bjørnøya). There, an undeformed sedimentary sequence, unconformably overlying the substrate, was found to consist of Upper Devonian sediments that showed a gradual transition up into layers corresponding to the Lower Carboniferous of Spitsbergen. Vogt (1928) took this to indicate that the post-folding unconformity at Spitsbergen was established prior to the deposition of the Bear Island sequence. This gave a <u>minimum age</u> of lower Upper Devonian, leading to the above age estimate for the Svalbardian folding. With respect to Scotland, Vogt (op.cit) pointed to a similar development there, with folded Middle Devonian layers unconformably overlain by Upper Devonian strata. Modern research on Svalbard has maintained that the Svalbardian orogeny occurred in the Upper Devonian (e.g. Chorowicz 1992; Dallmann 1992; Manby & Lyberis 1992). (See Sect. 2.9 for a review of the Solundian model).

Sturt (1983) was the first to use the name "<u>Solundian Orogeny</u>" for the E-W folding , etc., of western Norway. This new name implied a renaming of the traditional "Svalbardian Orogeny". Subsequently, the Solundian orogeny model was further emploid in several publications (e.g. Roberts 2003; Torsvik et al. 1986, 1987, 1988; Sturt & Braathen 2001).

<u>Model no. 4</u>: Transtension related to sinistral movements along the Møre–Trøndelag Fault Complex (MTFC) (External forces):

Sinistral movements along the MTFC, combined with a regional picture of transtension in western Norway, have been believed to produce N-S contraction during NW- and W-ward extensional movements in western Norway, leading to the folding of the Devonian basins and the rest of the Upper Plate into E-W trending folds (e.g. Krabbendam & Dewey 1998; Osmundsen et al. 1998; Osmundsen & Andersen 2001; Dewey & Strachan 2003). It should be noted that the model implies that the N-S contraction (= folding) in western Norway occurred *simultaneously* with the NW-SE or E-W extension in the area. The strain situation thoughout western Norway was found to be of overall constrictional type, consistent with this N-S contraction and E-W extension. As indicated by the above references, this model has been applied by several workers.

Model no. 5: Transpression or contraction, involving sinistral movements along the Møre–Trøndelag Fault Complex (MTFC) (External forces):

This model, which was detailed in Braathen (1999), sees the folding in western Norway as a result of a large-scale N-S shortening event. In Braathen (op.cit.), this event was illustrated as a result of the combined relative southward movement of the landmass located north of the Møre–Trøndelag Fault Complex at the time, and relative northward movement of southern Norway (see Figure 10 of Braathen 1999). The movements may have resulted in sinistral transpression along the Møre–Trøndelag Fault Complex. According to Braathen (1999), the observations “lend support to the suggestion that a **Late Devonian,** regional contractional or transpressional event post-dating Early Devonian extension (e.g. Roberts 1983; Sturt 1983), affected southern Norway”. Here, Braathen (1999) was referring to the abstract of Sturt (1983), where the Solundian orogeny model was proposed for the first time. However, Braathen (1999) focused on the *data* indicating N-S contraction, and did not discuss the Solundian model as such. It should be noted that the model of Braathen (1999) implied that the N-S contraction occurred *subsequently* to the NW-SE or E-W extension that formed the Devonian basins.

Model no. 6: Contraction being related directly to extension (Internal extension-regime forces):

In this model, the folding has been seen as having been produced solely by the extensional processes of western Norway: Strong E-W extension has been accompanied by N-S contraction. (“Pulling a cotton-towel”). The model has not gained support as an explanation for the folding.

Model no. 7: Primary corrugations on the detachment zone (Internal extension-regime forces):

In this model, the folding of the Upper Plate and the Devonian basins are believed to be partly a result of extensional movements that occurred on top of large-scale corrugations on the detachment zone (Chauvet & Seranne 1994). The model has not been widely applied.

Model no. 8: Folding of Devonian deposits as a result of “intrabasinal” transpression during westward movement of the basin (Internal extension-regime forces):

The model is based on data from the northern margin of the Hornelen Devonian Massif (Larsen 2002b). The westward movement of the basin took place above the Nordfjord–Sogn Detachment Zone, and oblique movement of the basin towards the northern fault produced transpressional forces that led to folding. The model, which has been applied by Larsen (2002b), is described below in Sect. 2.8.5.4. (See also Sect. 2.8.4.2).

Summary of fold models

In summary, fold-model no 4 (transtension due to MTFZ) is the model that has been mostly used in recent publications. Model no. 3. (Solundian) and no. 5. (transpression/ contraction involving the MTFZ) has also been supported in several recent publications.

As indicated above (c.f. the “head lines” of the paragraphs describing the models), the fold models may be divided into two groups based on whether the contraction was due to “external forces” (models no. 1–5) or “internal extension-regime forces” (models no. 6–8). The external contractional forces originated ***outside*** the area undergoing E-W extension and N-S contraction, whereas the internal contractional forces were a result of processes ***within*** the area experiencing this extension and contraction.

2.8.5.2 FOLD MODELS GROUPED ON THE BASIS OF TECTONIC REGIME: CALEDONIAN CONTRACTION OR POST-CALEDONIAN EXTENSION

From the above models, it appears that two different tectonic regimes have been envisaged for western Norway during the folding: ***either*** a regime where the area was in a state of overall late-/post-Caledonian extension (albeit combined with folding), where the extension was caused by e.g. transtension related to the MTFC (model 4) or several intra-extension-regime processes (models 6–8) – ***or*** a regime where the area was in a state of overall latest Caledonian orogenic-type contraction, represented by e.g. the Solundian orogenic phase (model 3) or N-S contraction/transpression involving the MTFC (model no 5). It has long been argued that the folding of the Upper Plate occurred *during* extensional movements (e.g. Hossack 1984; Norton 1987; Chauvet & Seranne 1989; Chauvet & Seranne 1994), but the controlling factors for, and the nature of the folding has been debated.

2.8.5.3 FOLD-MODELS GROUPED ON THE BASIS OF STRAIN SYSTEMS: CONTRACTION OR CONSTRICTION

Of the 7 fold-models presented above, no. 1–6 are connected to external large-scale tectonic processes or orogenic events. However, as also pointed out by Larsen (2002b), the fold-models proposed for western Norway may also be grouped on the bases of the strain systems invoked. Two groups may be suggested: Either, the folds were caused by a) bulk contractional N-S shortening, as in models no 3 and 5 above (e.g. Roberts 1983; Braathen 1999; Sturt & Braathen 2001), or by

b) bulk constrictional N-S shortening and E-W extension, as in model no 4 above (e.g. Chauvet & Seranne 1989; Chauvet & Seranne 1994; Hartz & Andresen 1997; Krabbendam & Dewey 1998; Osmundsen et al. 1998).

2.8.5.4 FOLDING OF THE HORNELEN BASIN: A RESULT OF INTRA-BASINAL TRANSPRESSION DURING WESTWARD MOVEMENT OF THE BASIN

As mentioned above, Larsen (2002b) found that alternative b) “bulk constrictional N-S shortening and E-W extension” was the strain system that was present in the Lower Plate basement gneisses *north* of Hornelen. Regarding the Hornelen *basin*, detailed analyses led Larsen (2002b) to conclude that the folding in the northern margin areas of the basin could <u>not</u> be explained by regional contractional N-S shortening as suggested by Braathen (1999) and Sturt & Braathen (2001). Instead, Larsen (2002b) concluded that the folding was exclusively a result of intra-basinal contraction due to transpression towards the basin margin faults during westward basin movements. When considering the whole region, including the Hornelen and the substrate, Larsen (2002b) concluded that the presence of *constriction* in the basement gneisses north of Hornelen, combined with the westward *extrusion* of the basin, indicated bulk *constrictional strain* in the whole region, although the folding of Hornelen beds was <u>not</u> a result of this regional constriction.

2.8.5.5 <u>COMMENT</u>: FOLDING OF THE HORNELEN BASIN BY "INTRA-BASINAL" TRANSPRESSION DURING WESTWARD MOVEMENT OF THE BASIN, MAKES "SOLUNDIAN OROGENY"-TYPE CONTRACTION UNLIKELY

If the folding in western Norway was caused by a "Solundian Orogeny"-type contraction, all units in the area should have been affected, i.e. the deformation should have been "unit exceeding". According to Larsen (2002b), however, the Hornelen folding was not caused by such *external* orogenic contraction, but instead *internal* processes. Since there appears to be no "unit exceeding" deformation affecting both the Hornelen massif and the adjacent Lower Plate alike, it seems that a model of N-S contraction of "orogenic-type", i.e. Solundian type, is too simple to explain the data, at least in *these* parts of western Norway.

The timing of the folding has been uncertain, but from palaeomagnetic investigations, Torsvik et al. (1988) suggested that the folding of the Devonian deposits occurred in the **Late Devonian–Early Carboniferous**. This age estimate is disputed below. Anyway, as also noted by Larsen (2002b), the folding of the subjacent Nordfjord–Sogn Detachment Zone and the Lower Plate did not necessarily occur simultaneously with the folding of the Devonian rocks. In the Hornelen area, Larsen (2002b) suggested that the folds in the Hornelen basin were somewhat younger than the basement folds.

2.8.5.6: <u>COMMENT</u>: INDICATIONS THAT FOLDING OF THE DEVONIAN DEPOSITS OCCURRED DURING THE DEVONIAN PERIOD, NOT THROUGHOUT THE LOWER CARBONIFEROUS: A TENTATIVE REVISION OF CONCLUSIONS DRAWN IN PUBLICATIONS ON PALAEOMAGNETIC DATING

Although uncertainties exist regarding the timing of folding of the Devonian massifs, many recent publications have assigned the folding to the **"Late Devonian–Early Carboniferous"** age (e.g. Braathen 1999; Osmundsen & Andersen 2001; Larsen 2002b), based on the age provided by Torsvik et al (1988) from palaeomagnetic analyses. However, as shown in the following, a re-examination of the paper of Torsvik et al. (1988) might indicate that this age estimate should be revised:

Torsvik et al. (1988, text p 427) stated that (quote)

> "*The relative pole-position obtained from the Devonian rocks of Kvamshesten, Håsteinen and Hornelen ORS (A components) and their substrates form a polar group around 15°N, 150°E (Fig. 18a), suggesting a* ***late Devonian/ early Carboniferous age****. These data are thought to reflect magnetic resetting during the crustal uplift history of the Svalbardian (Solundian) Orogeny*" (end quote).

(Bold types by present author). This "figure 18" of Torsvik et al. (1988) shows a plot of the *recorded* pole positions of Kvamshesten, Håsteinen and Hornelen Devonian Massifs (**Fig. 2.9a**). According to Torsvik et al. (1988, their text to figure 18), these pole positions are (quote)

> "*displayed on a provisional apparent polar wander path [APWP] given in Pesonen et al. (1987); c.f. also Sturt & Torsvik (1987), with A_{95} [= 95 %] confidence limits*" (end quote),

and, furthermore, the figure also shows (quote)

> "*the mean Devonian, Carboniferous, and Permian pole positions obtained from Fennoscandian data*" (end quote).

These established *mean* pole positions were compared with the *recorded* pole positions obtained from the three Devonian massifs. From Torsvik et al. (1988, their figure 18) (**Fig. 2.9a**), it appears that the *recorded poles* in all three massifs are situated well inside the **Devonian** pole position field, which have pole-positions around **15°N**, **150°E**, as stated by Torsvik et al (1988, p 427).

Torsvik et al. (1988) carried out a fold test on the data from Hornelen, and found that the magnetic vectors showed a minimum spread at **20%** unfolding, implying a late syntectonic (= synfolding) age for the pole position. In geological terms, this means that **80%** of the folding of the rock was accomplished before the magnetisation was "frozen" in the rock. The last **20%** of the folding thus led to a spread in orientation of the palaeomagetic vectors in the fold limbs. The **20%** *unfolding* in the foldtest, resulted in the removal of the same spread of the vectors.

Figure 18 of Torsvik et al. (1988) (**Fig. 2.9a**) shows that the palaeomagnetic data from the three massifs clearly fall within the error limits of the **Devonian** pole position, which is well away from the Carboniferous pole position. It therefore appears to be unfortunate that Torsvik et al. (1988) concluded that the folding occurred in the **Late Devonian–Early Carboniferous**.

As mentioned above, the APWP shown on figure 18 of Torsvik et al. (1988) (**Fig. 2.9.a**), was based on the APWP of Pesonen et al. (1987). However, a revised APWP for Baltica was subsequently presented by Torsvik et al. (1996). On an illustration of this 1996-APWP (**Fig. 2.9b**), the palaeomagnetic poles of Kvamshesten, Håsteinen and Hornelen (= "Kva-Hå-Ho") which are shown in Torsvik et al. (1988), have been plotted again (this work, in co-operation with Harald Walderhaug, Department of Earth Science, University of Bergen), with the same dp/dm confidence ovals, i.e. "error bars" as in Torsvik et al. (1988). ■ *From* ***Fig. 2.9b*** *it appears that the* <u>*mean*</u> *of the three poles fall closest to the* ***Late Devonian*** *time (c 362 Ma).* (The Devonian period span the interval **417–354 Ma**, i.e.

63 Ma, according to the time-scale of Gradstein & Ogg 1996, or the interval **416–359 Ma**, i.e **57 Ma**, according to the time-scale of Gradstein et al. 2004).

However, the almost “circular shape” of the Devonian segment of the APWP (**Fig. 2.9b**), around the Kva-Hå-Ho mean pole, implies that the mean pole is situated fairly close to *any* part of the Devonian APWP segment. ■ *This implies that the mean pole of the Kvamshesten, Håsteinen and Hornelen massifs could equally well fit any age within the Devonian period.* (Interpretation supported by Harald Walderhaug, pers. com. 2008). Accordingly, the mean pole would for example fit reasonably well with folding also in the **Early** or **Middle Devonian times**, in additon to the **Late Devonian**.

The formation of the Devonian basins of western Norway is dated to Middle Devonian on the basis of fossils, except in the Bulandet/Værlandet (Solund basin) where fossils appear to have a Lower Devonian age. The age of basin formation gives a maximun age for the folding. ■ *The palaeomagnetic data thus opens for the possibility that the basins were folded during extensional movements in **Middle or Late Devonian times**.*

The Kvamshesten and Håsteinen massifs were subjected to palaeomagnetic studies *prior* to the above referred investigation of Hornelen. From the investigation of <u>Kvamshesten</u>, Torsvik et al. (1986) concluded that the Solundian *folding* of the massif occurred in the **latest Devonian** (p 346), although the *magnetisation* was estimated to have taken place in the **Mid-Late Devonian to Early Carboniferous** (p 346). The fold test of the massif showed that the minimum spread of the magnetic vectors was obtained at **30%** unfolding, showing a late syntectonic (= late synfolding) palaeomagnetic pole, i.e. a situation equivalent to the one in Hornelen. Likewise, from the investigation of the <u>Håsteinen</u> massif, Torsvik et al (1987) concluded that the *folding* occurred in the **Late Devonian** (p 321), and the *magnetisation* apparently in the **latest Devonian** times (p 305). The fold test of Håsteinen gave the same result as for Kvamshesten; minimum vector spread was obtained at **30%** unfolding, again suggesting a late syntectonic (= late synfolding) pole, again equivalent to Hornelen.

As shown in **Fig. 2.9.b**, where the APWP is reproduced from Torsvik et al. (1996), the Kva-Hå-Ho poles cluster largely within the Devonian field. ■ *Hense, an important observation from Fig. 2.9.b is that the palaeomagnetic data give little reason to strech the pole age into the Carboniferous, to any great extent.* (Interpretation supported by Harald Walderhaug, pers. com. 2008).

As much as **80%** of the Hornelen folding, and **70%** of the Kvamshesten and Håsteinen folding, were completed largely within the Devonian period, i.e. before the magnetic vectors were “frozen” in the rocks. Larsen (2002b) concluded that the Hornelen basin was folded as a result of intrabasinal transpression during westward movement of the basin, i.e. during the post/late Caledonian regional extensional processes in western Norway. ■ *As this folding required the extensional processes to be active, it is very likely that the remaining 20–30% of the folding occurred soon after the magnetisation of the rocks, i.e. essentially within the **Devonian** period (possibly continuing slightly into the Lower Carboniferous).*

Fig. 2.9b shows that the pole for the Kvamshesten massif falls exactly on the **362 Ma** position on the APWP. ■ *This suggests that the magnetisation for this massif was aquired at about this time.* (Interpretation supported by Harald Walderhaug, pers. com 2008). It also implies that **70%** of the folding of the Kvamshesten massif was completed by the latest Devonian times. It is likely that the remaining **30%** of the folding occurred soon after, i.e. during the latest Devonian or earliest Carboniferous times.

As argued above, published palaeomagnetic and geological data appear to suggest that the folding of the Devonian rocks of western Norway occurred essentially during the Devonian. ■ *Accordingly, the proposal of Torsvik et al. (1988) that the folding also occurred throughout the Lower Carboniferous, which lasted **354–327 Ma**, i.e. **27 Ma** (time-scale of Gradstein & Ogg 1996), or **359–326 Ma**, i.e. **33 Ma** (timescale of Gradstein et al 2004), apparently cannot be inferred from the palaeomagnetic evidence, as is also, in fact, illustrated on Fig 18 of Torsvik et al (1988).*

In the earlier publication on Kvamshesten (Torsvik et al. 1986), the folding of this basin was suggested to have occurred in the Latest Devonian; and in the publication on Håsteinen (Torsvik et al. 1987), the folding of *this* basin was estimated to Late Devonian. On the 1996-APWP, these two age estimates lie within the displayed palaeomagnetic pole cluster. The Hornelen paper of Torsvik et al (1988) gives a synthesis on the folding of all the three Devonian basins. ■ ***In this paper, where the Kva-Hå-Ho poles lie entirely within the Devonian part of the APWP, no particular reasons were given for extending the folding-age into the Carboniferous, apart from the statement that the folding was caused by the Svalbardian/ Solundian orogenic phase, which they assigned to the Late Devonian–Early Carboniferous.***

From **Fig. 2.9b** it may be seen that the confidence ovals of the poles partly touch or cross the Devonian/Carboniferous boundary, which is set to **354 Ma** by Gradstein & Ogg (1996) and **359 Ma** by Gradstein et al. (2004). Hense, depending on the time-scale applied, or on future revisions of the APWP, it is possible that the confidence ovals could strech just into the Lower Carboniferous. However, as indicated above, the palaeomagnetic data of Torsvik et al. (1988; 1996) tends to support the argument that the folding took place largely within the **Devonian** period, and that it might actually have taken place at <u>*any time*</u> during this period.

In the literature dealing with folding of the Devonian deposits of western and central Norway, a large variety of models have been suggested to explain the folding, and several models do not associate the folding with the Solundian orogenic phase. Generally, since the *timing* of the folding imposes significant constraints on the modelling of the folding, it is important that this timing be as correct as possible. In the light of the above discussions, it appears appropriate to revise the widely referred age-interpretations of Torsvik et al. (1988), which states that the folding occurred in the Late Devonian and Early Carboniferous.

Hense, the data in Torsvik et al. (1988; 1996), might be reinterpreted to show that the folding was a **Devonian** event, i.e. not an event that also continued throughout the **Early Carboniferous**.

2.8.6 TWO MODES OF EXTENSION

Milnes et al. (1988, 1997) investigated the structural development along the northern shoreline of Sognefjorden, an area comprising Lower Plate rocks of the Western Gneiss Region (**Fig. 2.7**). The authors suggested *two modes* of extensional movement in the area. Three post-eclogite deformation phases (D_1–D_3) were recorded, and the area was divided into three structural regimes based on the intensity of this deformation, which was found to increase towards the Nordfjord–Sogn Detachment Zone in the west. D_1-shear foliations, and closely related D_2-folds with westerly vergence were related to a *first mode* of westward "back-thrusting" (Milnes et al. 1988) or "Mode-I extension" (Milnes et al. 1997, referring to Fossen 1992, see below), which occurred along the zone that they termed a *thrust* zone and which they called the Jotun Detachment (**Fig. 2.7**). Their Jotun Detachment had previously been established as the décollement for the Caledonian *thrust* emplacement of the Jotun Nappe. The subsequently formed Nordfjord–Sogn Detachment Zone, representing the *second mode* of extension (Milnes et al. 1988), or "Mode-II extension" (Milnes et al. 1997, referring to Fossen 1992, see below), was interpreted to cut the Jotun Detachment after the D_2-phase.

To the south and east of Sognefjorden, Fossen (1992, 1993a documented top-to-the-west shear criteria in the thrust décollement zone in the entire area between the Caledonian nappe front to the southeast, and the coast in the west (**Fig. 2.1**). Fossen (1992) were the first to relate these features to extensional westward movements of the entire Caledonian orogenic wedge of southern Norway. Two modes of extension were suggested for the whole area (Fossen1992); one early phase of westward back-movement along the old Caledonian thrust décollement, a movement which was termed "Mode-I extension"; and a later phase, called "Mode-II extension", with development of, and movement along detachment zones that cut through the décollement zone, notably the Hardangerfjord Shear Zone (Fossen & Hurich 2005) , which caused the downfalting of the Faltungsgraben rocks (**Fig. 2.2**); and the Nordfjord–Sogn Detachment Zone (which has been described above) with its southern continuation, the Bergen Arc Shear Zone. Further details of the extensional movements south and east of Sognefjorden were reported from the area between Øygarden and the Jotun Nappe (Fossen & Rykkelid 1992b), from the Øygarden Complex (Rykkelid & Fossen 1992), and the Bergsdalen Nappes (Fossen 1993b, 1993c; Fossen & Holst 1995). Radiometric dating was performed to find the age of the extensional movements. $^{40}Ar/^{39}Ar$-muscovite dating in the Bergsdalen Nappes (Fossen & Dallmeyer 1998) yielded ages between **403–398 Ma**, suggested to date the initiation of the décollement-cutting detachment/shear zones. Fossen & Dunlap (1998) were the first to date the initiation of extensional movements on the décollement zone in southern Norway. By dating sheared rocks iont he area between Bergen and the eastern part of the Jotun Nappe, it was revealed that eastward *thrusting* along the décollement zone occurred in the time interval **415–408 Ma**, whereas westward *extensional movements* along the same zone occurred between **402–394 Ma**, i.e. indicating that the extension in southern Norway probably started around **402 Ma**. Dunlap & Fossen (1998) applied K-fsp thermochronology to samples collected between Bergen and the eastern part of the Jotun Nappe, and found rapid cooling of the area from about **400 Ma** and tectonic and thermal stability between **380–330 Ma**. Detailed investigations of the Bergen Arc Shear Zone (BASZ) (Fossen 1992; Wennberg & Milnes 1994; Wennberg 1996; Milnes & Wennberg 1997; Wennberg et al. 1998; Wennberg et al. 2001), which is a zone situated along the eastern margin of the Bergen Arcs, has confirmed the detachment character of the zone and supported the interpretation that the BASZ is the southern continuation of the Nordfjord–Sogn Detachment Zone. In a study of post-Caledonian structures of the Bergen area, Fossen (1998) complimented the "Mode-I" and "Mode-II" termonology associated with Caledonian extension, by defining later brittle faulting as "Mode-III" extension. In the outer part of Hardangerfjorden, Norton (1987) suggested that a detachment zone defines the eastern boundary of the Sunnhordland Batholite on the islands of Bømlo, Stord and Tysnes (**Fig. 2.2**), and called it the Sunnhordland Detachment. A structural study of the Hardangerfjord Shear Zone was carried out by Fossen & Hurich (2005). The zone, which is NW-dipping along the entire SW-NE length of Hardangerfjorden, continues laterally northeastwards as the Lærdal–Gjende Fault system, and southwestwards into the North Sea where it can be traced on deep seismic sections. The zone is a Mode-II type detachment zone that cut the Caledonian nappe pile in the same mammer as the Nordfjord–Sogn Detachment Zone. The Hardangerfjord Shear Zone is defined by a **5 km** thick ductile (plastic) mylonite zone, and the dip-slip displacement was estimated to **10–15 km** as judged from the downfaulted Caledonian nappes. Because such a modest dip-slip would be insufficient to develop a **5 km** thick mylonite package, the shear zone was interpreted to have originated as a Proterozoic structure that had repeatedly been reactivated, now during the Mode-II movements. Fossen (2000) reviewed the extensional tectonics of southern Norway, and concluded that the Mode-I extension along the décollement zone reflected divergent plate motions between Baltica and Laurentia. Fossen (2000) concluded that the term “postorogenic” should be used for the time interval succeeding initiation of Mode-I extension, as opposed to the term “synorogenic”, which should be confined to the Caledonian time interval prior to this extension.

2.8.7 "DRIVING FORCE" FOR THE EXTENSION

Several models have been suggested to find the mechanisms and the **driving force** for the "collapse" of the Caledonian orogen and the resulting extension. Norton (1986) and Seranne et al. (1989) assumed the driving force to be the isostatic instability (gravitational forces) developing in the doubled crust when the Scandian orogenic contraction had ceased. Andersen & Jamtveit (1990), Andersen et al. (1991) and Dewey et al. (1993) relate the extension to uplift caused by the release (removeal) of the thermal boundary layer in the lithospheric mantle (Dewey 1988, England & Houseman 1988, 1989). This was disputed by Milnes (1997), Koyi et al. (1999), and Milnes & Koyi (2000) who argued that the uplift of the WGR was the result of the uplift of a buoyant lithosphere which had <u>not</u> been removed. Seranne et al. (1991) were the first to suggest that large-scale sinistral wrench tectonics, particularly involving the Møre-Trøndelag fault complex and the large British faults (e.g Highland Boundary Fault), may have contributed to the extensional movements in Western Norway. This model has subsequently been further developed by several authors (e.g. Krabbendam & Dewey 1998; Osmundsen & Andersen 2001). Fossen (1992) and Fossen & Rykkelid (1992) explained the extensional movements by a *de facto* reversal in plate motions from convergent to divergent, arguing that such a reversal is necessary to explain the documented westward movement of the Caledonian nappe wedge also in the eastern part of the orogen. This interpretation was later supported by e.g. Wilks & Cuthbert (1994), Milnes et al. (1997), Rey et al (1997), and the model was further substansiated by Fossen (2000). Among recent contributions, Walsh et al. (2007) advocated that the Western Gneiss Region (WGR) was exhumed *en block*. This was found to require that the WGR was either detached from a larger subducted slab and moved upwards in the manner shown by Chemenda et al. (2001), or exhumed as a result of reversal of the Baltic and Laurentian plate motion from convergent to divergent.

2.8.8 DEVONIAN DETACHMENT ZONES ELSEWHERE IN THE CALEDONIDES

Central and northern parts of Norway contain post-collisional detachment zones corresponding to the Nordfjord–Sogn Deatchment Zone of western Norway. In the following, these northern zones are presented, from south to north:

Southeast of Trondheim, the Røragen Detachment is situated below the Røragen Devonian Massif (Norton 1987, Norton et al. 1987; Gee et al. 1994). In a review of the area between Trondheim and Hamarøy (northern Nordland), Braathen et al. (2002) covered all the detachment zones found in the area. The southernmost of these, the Høybakken detachment (HD) (Seranne 1992b; Braathen et al. 2000, 2001; Kendrick et al. 2004; Eide et al. 2005; Osmundsen et al. 2006) is located below the Devonian massifs of the Fosen area at the coast of Trøndelag. The direction of movement has been top-WSW (i.e. towards **250°**). The Kollstraumen detachment (KD) (Braathen et al. 2000, 2001; Nordgulen et al. 2002) follows along the southeastern margin of the Helgeland Nappe Complex, and the direction of movement has here been interpreted to be top-ENE (i.e. towards **070°**), i.e. opposite of the Høybakken detachment. The Nesna Shear zone (NSZ) (Eide et al. 2002; Osmundsen et al. 2003) is positioned along the eastern and northern basal contact of the Helgeland Nappe Complex. East of this nappe complex, two detachment zones have been identified (Braathen et al. 2002), i.e. the Gaukarelv Shear Zone (GSZ), and the Seve-Køli boundary shear zone (S-KSZ). Further north, the Virvassdalen sherar zone (VSZ) is present along the western margin of the Nasafjäll basement window, and the Randal shear zone (RSZ) is localized within the window area (Braathen et al. 2002). North of Nasafjäll, the Breivika shear zone (BSZ) (Braathen et al. 2002) crosses Saltfjorden; and the Sagfjord shear zone (SSZ) (Braathen et al. 2002) is located in the districts east and south of Hamarøy island, partly following along the norhernmost boundary of the Uppermost Allochthon of this area.

Studies of onshore-offshore relationships have revealed that the detachments in central and northern Norway played a role in the construction of the offshore areas (Osmundsen et al. 2002; Olesen et al. 2002; Skilbrei et al. 2002).

In Northern Norway, detachment zones have been reported from the Ofoten area (Fossen & Rykkelid 1992a; Rykkelid & Andresen 1994; Klein & Steltenpohl 1990; Barker et al. 2000; Steltenpohl et al. 2004).

Outside Scadinavia, Caledonidan collapse-related extensional tectonics, with or without accompanying wrench tectonics, is also reported to controle formation of the Devonian massifs of East Greenland, the Orcadian of Scotland–Orkneys–Shetland, and Svalbard. These areas are briefly reviewed below:

In East Greenland, which represents large parts of the Laurentian craton that collided with Baltica during the Scandian orogeney, extensive research over the last decades have revealed that large-scale extensional tectonics and related exhumation of HP/UHP eclogites played an important role in the structuring of the region (McClay et al. 1986; Surlyk 1990; Larsen & Bengaard 1991; Hartz & Andresen 1995, 1996; Andresen et al. 1998; Hartz, et al. 1997, 2000, 2001; White & Hodges 2002, 2003; McClelland & Gilotti 2003; Gilotti & McClelland 2005; Sartini-Rideout et al. 2006; Andresen et al. 2007; Hang & Gilotti 2007). Radiometric $^{40}Ar/^{39}Ar$ muscovite dating in the major eastward-dipping crustal-scale Fjord Region Detachment zone (FRD), by Hartz et al. (2000), yielded progressively younger cooling ages downward along the detachment zone (i.e.eastwards), from **c. 408 Ma** in upper parts to **c. 388 Ma** in lower parts, reflecting that the deepest rocks were the last to cool below the retention temperature, as would be expected. The maximum time differrence between the earliest and latest cooling ages were estimated to **25 Ma**, and the zone was interpreted to have a long-lived history. White & Hodges (2002) carried out $^{40}Ar/^{39}Ar$ dating indicating that at least two distinct episodes of extensional movements took place along the Fjord Region Detachment (FRD), an early phase **c. 425-423 Ma**, occuring during overall orogenic contraction, and a post-Caledonian phase of reactivation at **c. 414 to 380 Ma**. The detachment were reported to be active again at **c. 357 Ma**, i.e. during deposition of Devonian sediments. Their data were interpreted to indicate that, rather than beeing active continuously for **80 million years** as suggested in the literature, the FRD were active for shorter intervals over discrete time periods separated by as much as **60 million years**. Exhumation of HP and UHP rocks was studied by McClelland & Gilotti (2003) and Hang & Gilotti (2007). The former concluded that the HP rocks were exhumed sometimes after **405 Ma**., whereas the latter assumed that the UHP metamorphism had lasted until **360–350 Ma** (see below). The prolonged contractional phase of the Greenland Caledonides (compared to the Scandinavian Caledonides) was investigated by Higgins et al. (2004) and Gilotti et al. (2004), of whom the last authors performed dating of eclogites. Sm-Nd of three HP eclogites yielded the ages **401 +/- 2 Ma**, **402 +/- 9 Ma**, and **414 +/- 18 Ma**. Additional $^{206}Pb/^{238}U$ dating of the same HP samples gave **401 +/- 7 Ma**, **414 +/- 13 Ma**, and **393 +/- 10 Ma**. Dating of an UHP eclogite yielded an $^{206}Pb/^{238}U$ age of **360 +/- 5 Ma**, an age which was found to be remarkably much younger than the HP samples. Dating the same UHP sample with the Sm-Nd method yielded an even younger age of **342 +/- 6 Ma**. Hense, Gilotti et al. (2004) arrived at the conclusion that the contractional phase in the Greenland Caledonides continued throughout the Devonian and possibly even into the Carbonifeous period, which is much longer than in the Scandinavian Caledonides.

The timing of the Greenland extensional phase, notably the Payer Land Detachment Zone, NE Greenland, was studied by Gilotti & McClelland (2005). Dating of the HP rocks of the lower plate of the detachment zone gave ages around **405 Ma** ($^{206}Pb/^{238}U$ zircon). Hense, the development of the detachment, and the related extension that caused exhumation of eclogites in the East Greenland Caledonides, was interpreted to be younger than **405 Ma**.

Sartini-Rideout et al (2006) studied the dextral strike-slip shear zones in Danmarkshavn, NE Greenland. These zones are located just west of the large-scale, dextral NNW-SSE trending Germania Land Deformation Zone (GLDZ). (The sinistral NNE-SSW trending Storstrømmen Shear Zone, present further to the east, contributes to devide these parts of the Greenaland Caledonides into a western , central and eastern block). Dating ($^{206}Pb/^{238}U$ zircon) of three sets of pegmatites that are partly deformed by and partly cross cutting the Danmarkshavn shear zones led the authors to conclude that the zones were active by **375 Ma**, and that the shear movements continued until, and stopped, at ca. **340 Ma**. Shear along both the Danmarkshavn shear zones and the closely associated GLDZ occurred coeval with formation of UHP eclogites at **360 Ma** in the eastern block further to the east (Gilotti et al. 2004). The authors suggested that vertical extrusion along the GLDZ in the period **370–340 Ma** could have caused the initial stages of exhumation of eclogites in the central block as well as a following exhumation of the eastern block after the **360 Ma** UHP metamorphism, lifting the rocks to amphibolite facies conditions. A second fhase of exhumation was suggested to have taken place in the late Carboniferous.

Andrsen et al (2007) investigated the time relationships between contraction and extension in the Kejser Franz Josef Fjord area, NE Greenland. The study area was located in the hanging wall of the Fjord Region Detachement. Samples dated by the U-Pb method were collected from the pre-to *syn-contractional* rocks of the Krummedal sequence, and the post-contraction–*syn-extensional* Grejsdalen pluton (intruding the Eleonore Bay Supergroup). Syn-contractional magmatic rocks, reflecting mid- to deep crustal levels, yielded ages in the interval **429–425 Ma** (zircon, monazite, xenotime). A syn-extensional granite pluton, reflecting shallow crustal conditions, gave an age of **429 +/- 1 Ma**. From this the auhors concluded that subhorizontal upper crustal extension was *contemporaneous* with mid- to lower crustal subhorisontal contraction.

Smith et al. (2007) analysed the microstructural evolution of the thrust-related Imbricate Zone, and the sinistral strike-slip Storstrømmen Shear Zone (SSZ), in the Hertugen af Orleans Land and Dronning Louise Land, NE Greenland. The NNE-SSW trending Imbricate Zone, situated west of the SSZ; marks the western limit of the area affected by Caledonian contraction, whereas the SSZ transects the orogenic hinterland. Based on published $^{40}Ar/^{39}Ar$ ages (Dallmeyer et al. 1994), Caledonian contraction in the Imbricate Zone was estimated to have taken place in the interval **400–380 Ma**. Based on several referred, published $^{40}Ar/^{39}Ar$ muscovite and U-Pb zircon datings, the ductile shear movements on the Imbricate Zone was estimated to have occurred in the period **390–370 Ma**. Hense, the authors advocated that the two zones were active brodly contemporaneously. The movements were also found to have taken place at the same structural levels. According to the authors, the conclusions of the study "support models that involve sinistrally oblique Silurian–Devonian collision between Laurentia and Baltica", referring to Soper et al. (1992) and Dewey & Strachan (2003).

Hang & Gilotti (2007) studied partial melting of metapelites at UHP conditions at the Rabbit Ears Island at **78°N** in the bay of Jøkelbugt, NE Greenland. UHP eclogites with confirmed coesite were reported to be present **<1.5 km** away from the metapelites. Although coesite was reported to be presently absent in the metapelite, the mineral was believed to have been present during the UHP conditions. The UHP rocks formed at pressure-temperature conditions of around **970°C** and **3.6 GPa**, referring to Gilotti & Ravna (2002), at an age of around **360 Ma**, referring to Gilotti et al. (2004).

Extensional zones of corresponding type; or transcurrent faults; have also been found to controle Devonian sedimentation in the <u>Orcadian Basin of Scotland–Orkeys–Shetland</u> (McClay et al. 1986; Norton et al. 1987; Seranne et al.1991; Hillier & Marshall 1992; Seranne 1992a; Watts et al. 2007). — and on <u>Svalbard</u> (Manby & Lyberis 1992; Chorowicz, 1992; McCann & Dallmann 1996; Friend et al. 1997; Blomeier et al. 2003).

2.8.9 SUMMARY

As illustrated above, the "detachment" model in western Norway integrates (i) data on eclogite-formation as well as uplift-related retrogression, (ii) the structural and metamorphic development of the mylonite belt (the detachment zone), (iii) the juxtaposition of greenschist facies and eclogite facies rocks, and (iv) the formation, sedimentation, deformation and metamorphism of the Devonian Basins, into one uniform model. The "detachment" model is the the first model that has succeeded in relating these data in space and time, and, hence, has finally provided an elegant explanation for several of the most important and problematic geological features in these parts of western Norway, features which could not earlier be explained.

2.9 THE "SOLUNDIAN OROGENY" MODEL

2.9.1 ESTABLISHMENT OF THE MODEL

Simultaneously with the development of the "detachment" model during the 1980s – which stated that the Caledonian Orogeny had ceased at the onset of large-scale extensional tectonics and widespread formation of Devonian basins – the opposite view was taken by Sturt (1983; 1984) and co-workers, who stated that the Caledonian contraction continued after formation of the Devonian basins, arguing that both the Devonian massifs and the whole area between Hardangerfjorden and Trondheimsfjorden had been involved in a latest Caledonian orogenic phase called the ***Solundian orogeny***. Their conclusions lent renewed support to the traditional school of thought stating that the deformation of the Devonian massifs was due to a Svalbardian orogenic phase (Vogt 1928; Roberts 1983).

The views held by Sturt and co-workers on the Solundian Orogeny, were developed in two seperate stages of research. This resultet in two different "Solundian" models, deviating from each other in terms of what geological features that were assigned to the Orogeny. Publications from the first stage of research appeared in the years 1983-1988, whilst publications of the second phase appeared a decade later, in 1999-2001.

The model used in the first phase of research will here be termed "Version 1", whilst the model from the second phase is denoted "Version 2". In essence, the difference between the two versions is that in "Version 1" the detachment mylonites below the Devonian massifs were interpreted as related to ***eastward thrusting*** of the massifs and the Upper Plate, whereas in "Version 2" the mylonites are seen as related to ***westward extensional movements*** of the units. The "thrust"-interpretation in Version 1 implied that the abundant top-to-the-west, intra-mylonite-zone shear sense indicators, which were well documented in the literature at the time, were ignored. This was corrected in Version 2, where the "extension"-interpretation implied that the shear sense indicators were acknowleged. The following paragraphs give a short outline of the two versions of the Solundian orogeny model.

2.9.2 VERSION 1 OF THE SOLUNDIAN OROGENY MODEL

In Version 1 of the Solundian Orogeney model, the arguments were based on tectonometamorphic, tectonomagnetic and palaeomagnetic data from the Devonian massifs and substrates (Bøe et al. 1989; Torsvik et al. 1986, 1987, 1988; Ramsay et al. 1987).

The geological phenomena explained by their model were the following (Torsvik et al. 1988):

- E-W trending gentle folds
- axial plane cleavage that formed in the E-W folds
- lineations: pebble lineations in the Devonain, stretching lineations in the substrate
- magnetic fabrics
- brittle faults: steep and low-angle, located along the margins of the Devonian Massifs.

The presence of the nappe pile formed by Scandian crustal-scale contractional deformation and imbrication formed the starting point for the Solundian model. The main conclusions in the model were as follows:

- the deposition of the ORS may well have been a result of extensional tectonics. The deformation and associated metamorphism, however, occurred subsequent to and independent of the depositional phase, and must be viewed separately (Torsvik et al.1988).
- the mylonite zone below the Kvamshesten Devonian Massif was interpreted as related to eastward thrusting during the Late Devonian Solundian (Svalbardian) Orogeny (Torsvik et al. 1996). By implication, this interpretation would also apply to mylonite zones below the *other* Devonian massifs in western Norway. This Solundian orogeny was interpreted as a continuation of the major Scandian phase of the Caledonian Orogeny (Torsvik et al. 1986).
- deformational features in Devonian rocks (E-W trending folds, axial plane cleavages, stretching lineations and pebble lineations, "illite crystallinity" and magnetic properties) were caused by N-S compressional forces related to the Solundian (Svalbardian) Orogeny (Torsvik et al. 1988).
- large-scale folds to the east of the Devonian Massifs, i.e. to the east of the Hornelen Devonian Massif, and also, for instance, the Surnadalen Syncline, as well as folding of the thrust plane below the Kvamshesten Devonian Massif, were a result of Solundian orogenic type compression. The Solundian Orogeny was claimed to have affected the rocks from the Hardangerfjord to the Trondheimsfjord (Sturt 1983).
- the present margins along the Devonian massifs are defined by faults related to Mesozoic graben formation (e.g. in the Hornelen Devonian Massif, Torsvik et al. 1988), and were not the sites of Devonian lateral ramps nor oblique slip in an extensional collapse model (Torsvik et al. 1988).
- geothermometry/-barometry based on illite crystallinity and phengite compositions from Devonian rocks have yielded temperatures of metamorphism exceeding **300°C** for the Hornelen Devonian Massif (Torsvik et al. 1988) and for the Devonian rocks of Smøla (Bøe et al. 1989), and pressures in the range of **4.1– 4.8 Kbar** for the Devonian rocks of Smøla, (Bøe et al. 1989). Such high temperatures and pressures must be due to orogenic dynamothermal processes, probably due to overriding nappes, i.e. the Solundian Orogeny.

It should be noted that the overall geologic setting in western Norway was not the subject of their study, and therefore the juxtaposition of, for instance, the eclogite terrane and the lower greenschist facies terrane was not considered. The study also did not attempt to shed any new light on the *formation* of the Devonian massifs (Torsvik et al. 1988).

Hense, in their Version 1 of the orogeny model they did not give Devonian extension and large-scale detachments any important role in the shaping the rock arcitecture of western Norway.

2.9.3 VERSION 2 OF THE SOLUNDIAN OROGENY MODEL

2.9.3.1 MAJOR ELEMENTS OF THE MODEL. —WITH CRITICAL AND SUPPLEMENTING COMMENTS.

Version 2 of the Solundian Orogeny model was mainly based on tectonometamorphic and structural data aquired by Sturt & Braathen (1999 report; 2001paper) from the Solund Devonian Massif. In their published paper, Sturt & Braathen (2001) frequently referred to the analyses of Braathen (1999), who interpreted the E-W trending folds in western Norway, and the S-directed thrusting in Kvamshesten, as related to regional N-S contraction. However, since Braathen (1999) did not explicitly relate these features to a Solundian Orogeny, the brief review of "Version 2", following below, will be based on Sturt & Braathen (2001). These authors strongly argued that the Solundian orogeny was responsible for the folding and other types of deformation of the Solund Devonian Massif, and also for the E-W trending folds of western Norway and the NE-SW trending folds of Trondheimsleden (Sturt & Braathen 2001, p 281 and p 283).

During the 1990s, numerous publications (e.g. Andersen & Jamtveit 1990; Swensson & Andersen 1991; Dewey et al. 1993; Wilks & Cuthbert 1994; Krabbendam & Dewey 1997; Andersen 1998) continued to document the fundamental role played by the mylonitic top-to-the-west extensional Nordfjord–Sogn Detachment Zone in the shaping of the rock architecture of western Norway, notably its role in the juxtaposition of Lower Plate eclogites against Upper Plate greenscist-facies rocks. As mentioned above, Sturt and Braathen (2001), in Version 2, therefore had to revise their Version 1 of the Solundain orogeny model. In Version 2 of the model, the crustal-scale extension was appreciated as beeing fully responsible for the construction of the overall rock architecture of western Norway. In particular, the mylonites below the Devonian massifs were now accepted as defining the Nordfjord–Sogn Detachment Zone, implying that the mylonites were accepted as having been formed during top-to-the-west/northwest *extension*, and not during top-to-the-southeast *thrusting* as argued in Version 1 of the model.

Although the detachment mylonites had now been abandoned as evidene for Solundian contraction, Version 2 maintained that the folding and other contractional structures of the Devonian basins were related to a phase of N-S regional contraction that was believed to have taken place *subsequent* to the extensional tectonics and the formation of Devonian basins. The following geological features of the Solund Devonian Massif were now taken to favour Solundian contraction (Sturt & Braathen 2001. Page numbers given below refer to this paper):

- Thrust-emplaced basement-cover slices (p 272): Three slices of rocks of substrate-type, found within Devonian sediments at Lifjell in eastern part of the Solund massif, were reported to have basal contacts defined by 2 m thick semiductile mylonite zones, and to have upper contacts being primary unconformities. The basal mylonites were reported to show top-to-the-SE movement, and the slices were interpreted as thrust repetitions of basement-cover.

Comment: The slices appear to lie parallel to bedding. This could indicate that the slices are landslide bodies, analogous to similar bodies at central Solund. However, this possible alternative interpretation was not discussed. Southeastward thrusting of the slices, as suggested by Sturt & Braathen (2001), would imply that the slices have been thrusted southeastwards across the Devonian rocks lying northwest of the bodies.

- Deformation structures at Utvær (p 274): At these westernmost Solund islands, intense deformation of Devonian bedding was reported to have produced structures like folds, transposed bedding, strongly penetrative cleavage being variably axial planar to the folds, and a strong degree of rotation of pebbles into the cleavage. The unconformity was reported to be tightly folded, and in places overturned. It was refered to Indrevær & Steel (1975), who had earlier related the deformation structures to a "thrust at depth" reported by Nilsen (1968).

  Comment: The Utvær folds have NW-SE trends, i.e. oblique to the E-W trending folds elsewhere in Western Norway. This deviating fold orientation was not discussed by Sturt & Braathen (2001). In the Solund massif, the Lågøy anticline is developed across the entire massif from the NE to the SW. The hinge line of the anticline has a curved trend, with a NE-SW trend in the NE, and a NNE-SSW trend in the SW (Nilsen 1968). In the latter area, the crest of the anticline passes near the NW-SE trending Utvær folds, which thus appear to make an angle of **45°** with the trend of the Lågøy anticline. This apparent difference in trend was not discussed by the authors.

- Lower greenschist-facies metamorphism (p 279): The tectonometamorphic fabric was reported to be semi-schistose in nature, with a cleavage mainly defined by chlorite and some epidote. Other characteristic minerals were reported to be recrystallised quartz and epidote, with sporadic small biotite porphyroblasts. This mineral paragenesis was interpreted as indicative of lower greenschist-facies metamorphism. The authors (Sturt & Braathen 2001) found this to be consistent with Svensen & Jamtveit (1999 abstract), who had, according to Sturt & Braathen (2001): (i) reported presence of "biotite and amphibole (unspecified) in addition to quartz and calcite" in veins cutting the sediments of the Solund massif, and who had also (ii) generally stated that growth of chlorite and epidote was indicative of greenschist-facies metamorphism.

  Supplementing information from Svensen et al. (2001): In the *abstract* of Svensen & Jamtveit (1999), which gave preliminary results from their study of vein inclusions and authigenic minerals of Solund and other Devonian basins, the presence of biotite was taken to indicate temperatures in excess of **~300°C**. The subsequent *paper* by Svensen et al. (2001) gave the full report from their study. This work confirmed that the temperature during maximum burrial of the Solund basin had exceeded **~300°C**, and in addition it was reported that chlorite thermometry had yielded a temperature in the range **305–330°C**. The burrial of the Solund sediments was estimated to **13 km**.

  Comment: The estimated burrial depth of **13 km** in Svensen et al. (2001) was possibly obtained by combining a thermal gradient of e.g. **25°C/km** with the temperature of **330°C**, giving the relationship **330°C / 25°C/km = 13.2 km**.

- Geological features discrediting other models (p 280): The following features were taken to discredit models for synchronous regional extension and N-S shortening: i) The lack of unconformities within the Devonian deposits, ii) the presence of tight folds in the Devonian, and overturned bedding and unconformity, and iii) the combination of (i) ***deformation/metamorphism*** of the Devonian rocks, coeval with the (ii) ***deposition*** of the same rocks, which was taken to be an impossible combination.
- Deformation style indicating compressional deformation (p 281): According to the authors, who also referred to Roberts (1983), the tectonic structures were of a style which one normally associated with a compressional, or at most transpressional, mode of deformation, and not one which could be expected to develop in an extensional regime. Structures particularly attesting to this were i) deformation features like overturned basement-cover contacts, ii) transposition of primary features into the cleavage, iii) pebbel rotation, etc. According to the authors, the folding and cleavage development had been accompanied or succeeded by strong flattening deformation, at least locally.
- Tectonothermal development was reminiscent of that at Hitra/Smøla (p 282): The metamorphism and deformation of the Solund Devonian rocks of Utvær were taken to be indicative of a compressional deformational regime at intermediate crustal depth. According to the authors, this was reminiscent of the tectonothermal development of the Devonian rocks of Hitra and Smøla, where Bøe et al. (1989) carried out P-T analyses on phengite, yilding a pressure in the range **4.1– 4.8 Kbar** and a temperature in the range **300–350°C**.

2.9.3.2 ORIGIN OF THE CONTRACTIONAL FORCES.

In Version 2 of the "Solundian orogeny", two different models were presented by Sturt & Braathen (2001) to account for the regional contractional processes that deformed the Devonian rocks:

i) Collision between an island arc and Baltica in the Late Devonian–Early Carboniferous:

In this model, proposed for the first time by Sturt & Braathen (2001), the Solundian phase was envisaged as resulting from a **Late Devonian–Early Carboniferous** collision between an island arc, sited near Shetland, and the continental block of Baltica. In order to find support for this, Sturt & Braathen (2001) referred to Mykura (1976), Thirlwall (1981, 1983 reply) and Astin (1983 discussion) who dealt with volcanic rocks within the sedimentary Middle Devonian deposits of Shetland and Orkney. Particular emphasis was on Thirlwall (1981), who had reported that these extrusive rocks were of calc-alkaline bimodal composition, which indicated that the rocks were subduction-related island arc volcanics.

Comment: Indications that the proposed collision between an island arc and Baltica in the ***Late Devonian–Early Carboniferous*** *did not take place:* The proposal by Sturt & Braathen (2001) of an island arc–Baltica collision in the Late Devonian–Early Carboniferous is problematic. Seven years after the appearance of the paper of Thirlwall (1981), Thirlwall (1988) concluded that "the best evidence for the timing of closure of Iapetus is provided by post-Downtonian (**post-412 Ma**) and **pre-394 Ma** deformation in the Lake District, and post Lower Old Red Sandstone (**post-410 Ma**) deformation in the Midland Valley". In other words, Thirwall (1988) concluded that Iapetus closed prior to **394 Ma** (upper Early Devonian, time-scale of Gradstein & Ogg 1996; lower Middle Devonian, time-scale of Gradstein et al. 2004). The suggestion by Sturt & Braathen (2001) that the island arc collision had occurred in Late Devonian or later, implies that the collision took place subsequent to **370 Ma** (time-scale of Gradstein & Ogg 1996) (or subsequent **385 Ma**, time-scale of Gradstein et al. 2004). This would be at least **24 million years** (or **15 million years**) after the closure of Iapetus at **394 Ma**. Hence, the conclusions of Thirlwall (1988) on the timing of the Iapetus closure, indicates that the "island arc–Baltica collision" model, proposed by Sturt & Braathen (2001), appear to be unrealistic.

ii) Transpression along the Møre-Trøndelag Fault Complex (MTFC):

In this model, regional N-S contraction of western Norway was suggested to be accompanied by sinistral transpression along the MTFC. The authors referred to Braathen (1999) for details on this model.

Comment: The folds in the Solund massif, i.e. the NW-SE-trending local folds at Utvær, and the NE-SW to NNE-SSW trending large-scale Lågøy anticline through the entire Solund massif, are all directed *obliquely* to the ~E-W trending folds found elsewhere in western Norway. Sturt & Braathen (2001) did not discuss the deviating fold trends of the Solund massif, but merely stated that the local folds at Utvær were parasittic folds to the large-scale folds.

2.9.3.3 EARLY CARBONIFEROUS UNROOFING IN SUNNFJORD

Eide et al. (1999) reported **Early Carboniferous** unroofing in the Sunnfjord region of western Norway, based on $^{40}Ar/^{39}Ar$ thermochronology on alkali feldspar (K-fsp). The study area was located west and south of the Kvamshesten Devonian Massif. The samples, which were taken from the Western Gneiss Region at the island of Bårdsholmen south of Askvoll, and from the Høyvik Group at Stongfjorden north of Askvoll, showed rapid cooling at **≥15°C/m.yr**. in the time interval **360–340 Ma**, i.e. in the **Late Devonian–Early Carboniferous**. (The Devonian/ Carboniferous boundary is at **354 Ma**, time-scale of Gradstein & Ogg 1996; and at **359 Ma**, time-scale of Gradstein et al. 2004). According to figure 5 in Eide et al. (1999), the crustal level, reflected by the WGR sample, showed cooling from **~285°C** to **~135°C** (i.e. a temperature-drop of **~150°C**) in the time interval **ca. 360–350 Ma**, i.e. during **~10 m.yr**. The Høyvik Group showed cooling from **~200°C** to **~115°C** (i.e. a temperature-drop of **~85°C**) in the interval **ca. 345–340 Ma**, i.e. during **~5 m.yr**. (The Devonian/Carboniferous bondary is at **354 Ma** or **359 Ma**, see above). This unroofing event was interpreted to be a result of tectonic and erosional processes, possibly induced by <u>thermal underplating</u>. It was suggested that the thermal underplating, if precent, <u>did not produce extensive melting and magma production</u>. Eide et al. (1999) estimated that the event occurred in the time interval **Late Devonian–Early Carboniferous**, an estimate which was found to be <u>in accordance with the timing of the Solundian Orogeny</u>, which, with reference to Torsvik et al (1986), was claimed to be responsible for the E-W folds, etc., in western Norway. Eide et al. (1999) drew attention to two published regional models that other workers have suggested could explain the unroofing, folding, etc., in this time period: (i) transcurrent sinistral motion along the

Laurentia–Baltica suture, possibly focused along the Møre–Trøndelag Fault complex, or (ii) “back-arc extension related to subduction of the Rheic ocean as the European Massifs and Gondwava advanced north toward Euramerica [producing the Variscan orogeny*; comment by the thesis author*] or as a response from the docking of Kazakhstan from the east”. Eide et al. (1999), however, pointed to the lack of strong evidence for any *specific* external tectonic event, and thus did not draw any conclusions on the matter.

Sturt & Braathen (2001), who proposed that the deformation of the Devonian rocks in western Norway could have been caused by a Late Devonian–Early Carboniferous collision between an island arc (situated near Shetland) and the Baltic continent, did not refer to the “thermal underplating” suggested by Eide et al. (1999), although the events would appear to coincide in time. Nevertheless, as argued above (Sect. 2.9.3.2), the proposed island arc–Baltica collision is probably unrealistic.

Absence of Early Carboniferous unroofing in the Bergen–Sogn–Bygdin region, but instead Lower Devonian and Permo-Carboniferous unroofing in this area

Fossen & Dunlap (1998) applied K-feldspar $^{40}Ar/^{39}Ar$ geochronology (i.e. the same method as Eide et al. 1999) to samples collected along a traverse across the Caledonides of south Norway. The route of the traverse, which had **7** sample localities, was from Bergen – along the NW boundary of the Jotun Nappe (Vikafjellet–Sogn–Sognefjellet) – to Bygdin at the eastern margin of the Jotun Nappe. The sample localities at the NW part of the traverse were positioned at least **80 km** southeast of the above-mentioned Bårdsholmen locality of Eide et al (1999). Fossen & Dunlap (1998) did not find the Lower Carboniferous (**360–340 Ma**) rapid unroofing event reported by Eide et al. (1999). On the contrary, Fossen & Dunlap (1998) found thermal and tectonic stability in the period **380–330 Ma** (or possibly **380–300 Ma**). Rapid cooling was, however, reported to have occurred (i) in the Early Devonian (**~ 400 Ma**), interpreted as related to onset of extensional collapse of the Caledonian mountain chain, and (ii) in the Permo–Carboniferous times (**330–250 Ma**), interpreted as related to onset of rifting in the Oslo Graben and the North Sea region.

2.9.4 SUMMARY AND EVALUATION OF THE “SOLUNDIAN OROGENY”-MODEL

As shown above, over the last decades, changes have occurred as to what deformational structures be ascribed to the “Solundian Orogeny”. In Version 1 of the Solundian model, the top-to-the-*west* mylonites of the Nordfjord–Sogn Detachment Zone were erroneously interpreted as top-to-the-*east* mylonites related to Solundian thrusting. This error was corrected in Version 2 of the model, as the mylonites were no longer part of the Solundian model.

Until recently, the folding of the Devonian basins and substrate into E-W trending folds has been considered by several authors to be a result of a regional phase of N-S contraction, i.e. the Solundian. However, the work of Larsen (2002b) suggested that the folding of the Hornelen Devonian basin was not caused by such a regional contraction, but by “intrabasinal” transpression during westward motion of the basin. This new interpretation suggests that the folding of the Devonian massifs of western Norway should be re-investigated to reveal whether other such folds may also be a result of “intra-basinal” transpression, and thus entitled to be “removed” from the “Solundian” portfolio of “regional” deformational features. In this picture, it could also be of interest to investigate whether the entire Upper Plate was deformed in a similar setting.

The unroofing described by Eide et al. (1999) was interpreted to indicate that the WGR experinced uplift in the Late Devonian–Early Carboniferous. Eide et al. (1999) referred to the statement of Osmundsen et al. (1998) that the postcollisional exhumation period was completed by the Early Carboniferous – and therefore advocated that the subsequent sudden “unroofing” during Early Carboniferous time must have had a different, but not necessarily unrelated, cause.

Late Devonian deformation and Early Carboniferous unroofing in the Sunnfjord Region of western Norway, is still poorly understood. Furthermore, as also pointed out by Eide et al. (1999), there is a lack of evidence for any “external” large-scale tectonic processes of “Solundian orogeny” type that could cause the deformation and uplift. It is important to note that, in the Jotun Nappe region further to the SE, and in the Bergen area to the S, Fossen & Dunlap (1998) found no signs of unroofing in this period. This apparent paradox could either indicate that the unroofing was confined to areas further to the NW and N, or it could indicate unappreciated problems with the methods or interpretations.

Generally, it cannot be excluded that many of the features assigned to the “Solundian”, have other explanations. At present, it still remains to be proven that a “Solundian Orogeny” exists.

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Chapter 3

CHAPTER 3 EIKEFJORD GROUP

3.1 INTRODUCTION

Chapter 3 deals with the northern to northeastern part of the study area, i.e the area to the north of the Sunnar Fault [(1)] (**Fig. 3.1**, **Plate 1**).

The rocks of the area are mainly orthogneisses which are generally extensively mylonitized, and this mylonite fabric is folded into east-west trending, mainly gentle to close folds. The mylonite fabric and related structures, and the folds (of post-shear or possibly latest syn-shear age), completely dominate the structural picture (**Figs. 3.2** and **3.3**). The rocks show a retrograde metamorphic development from amphibolite facies to greenschist facies.

The purpose of the investigation of these orthogneisses has been to provide tectonometamorphic data for a consideration of the extent to which the mylonites and the folds might be connected to the formation and deformation of the Håsteinen Devonian Massif, or in other words — whether the mylonites are part of the *Nordfjord–Sogn Detachment Zone* or not. This question is discussed briefly in Sect. 3.5, and more extensively in Ch. 6. The present chapter will mainly present the observed data from the area.

Data have mainly been recorded along two N-S trending road sections, providing profiles more or less orthogonal to the strike direction/trend of the structural elements in the rocks. The profiles investigated are (1) Eikefjord–Sunnarvik and (2) Osa–Vasset (**Appendix B: Road log no. 4**). In addition, investigations have also been carried out in the area located *between* the two profiles, an area bordered to the *north* by the E-W road section Sunnarvik–Osa, which provides exposures parallel to the stretching lineations and fold axes, and bordered to the *south* by the Sunnar Fault (**Plate 1**).

The chapter starts with an overview of some general features relevant to the Eikefjord Group (Sect. 3.2), and continues with rock descriptions (Sect. 3.3). Thereafter, the structural history of the group is treated (Sect. 3.4). The structural investigation has been confined to an analysis of the main structural elements of the mylonite fabric and the folds. This is followed by a presentation of the metamorphic development (Sect. 3.5), which is succeeded by a summary of data and interpretations for the Eikefjord Group (Sect. 3.6). The Sunnar Fault is presented at the end of the chapter (Sect. 3.7). The objective in the descriptions of the Eikefjord Group has been to establish and document the *overall* structural and metamorphic state of the rocks, and

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[(1)] Note that the name “Sunnar Fault” has been changed to “Sunnarvik Fault” on the 1:50 000 maps of Bryhni & Lutro 2000a, 2000b, a name also adopted by Johnston et a. 2007b. However, the original name “Sunnar Fault” will be used throughout the present thesis.

emphasis is therefore not on detailed discussions of single structural elements. For the purposes of the present investigation it was not necessary to carry out detailed mapping of the different rock types, hence, the Eikefjord Group has not been differentiated on the main map (**Plate 1**).

3.2 GENERAL FEATURES

The present sub-chapter contains a brief description of some important general features in common for the Eikefjord Group. This includes a discussion of • the naming of the group, • unit boundaries in the valley eastwards from the village of Eikefjord, • regional correlations, and • radiometric ages.

Naming

The designation "Eikefjord Group" was introduced by Bryhni et al. (1981) and also used later (e.g. Bryhni 1989; Bryhni & Lutro 1991a, 1991b, 2000a, 2000a; Wilks & Cuthbert 1994; Johnston et al. 2007a, 2007b) (see discussion in Ch. 2 on the use of this name). According to the rules of the Norwegian Committee on Stratigraphy (Nystuen 1989), the term "group" should have been replaced by the term "suite". However, since the present author has not carried out detailed studies in the Eikefjord Group as a whole, the name has not been changed in the present work.

When references are made to the Devonian deposits, the term *basin* means the original Devonian basin as it was at the time of deposition or soon after, whereas the term *massif* means the present-day remnants of the original Devonian basin. When referring from litterature that use the term *basin* in both cases, the term is used as in this literature.

Boundaries

The northern boundary of the Eikefjord Group (in the Eikefjord Valley) is defined by the E-W trending Eikefjord Fault (**Fig. 2.6**), running from the sea shore close to Eikefjord School (UTM 1230 3385) and eastwards (Bryhni et al. 1981). The southern boundary is constituted by the Sunnar Fault.

The Eikefjord Fault**:** In the Eikefjord area, the fault is defined by an up to **10 m** high fault escarpment where the northern block is the higher side. This western part of the fault separates the Eikefjord Group to the south from the Lykkjebø Group to the north (**Fig. 2.4**), and both these groups show the same penetrative mylonitic development. The fault cuts through these mylonites and juxtaposes the two groups. At Eikefjord School, the fault cliff contains a coherent fault breccia where fragments of the wall rocks are surrounded by a dark green chloritic groundmass. The amount and direction of displacement on the Eikefjord Fault are uncertain. Bryhni (1958) reported that meta-supracrustal rocks of the Lykkjebø Group rested on top of the meta-anorthositic rocks of the Eikefjord Group to the north and northwest of the village of Eikefjord. Also on map-figures in Andersen & Jamtveit (1990) and Furnes et al. (1990), similar tectonostratigraphic positions are inferred. The juxtaposition of the tectonostratigraphic higher unit (Lykkjebø group), situated north of the fault, with the tectonostratigraphic lower Eikefjord Group to the south, may suggest relative downward movement of the northern block, as was suggested by Bryhni (1964a). $^{40}Ar/^{39}Ar$ dating on white mica from both sides of the fault has been reported by Berry et al. (1993; 1995) and Andersen (1998), yielding **~ 415 Ma** on the northern side, and **~ 400 Ma** on the southern side, i.e. a difference of **~15 Ma**. The ages were taken to suggest that the northern part was uplifted above the Ar-retention temperature (of **~350 ºC**) **ca. 15 Ma** *earlier* than the southern part. However, such a conclusion is contradicted by a $^{40}Ar/^{39}Ar$ -muscovite dating at **403 +/-1 Ma** obtained by Chauvet & Dallmeyer (1992) from the Lykkjebø area, an area which is located between the fault and the **~ 415 Ma** locality on the *northern* side of the Eikefjord Fault, suggesting that the $^{40}Ar/^{39}Ar$ -muscovite ages are centred around **400 Ma** on *both* sides of the fault. Berry et al (1995) considered the Eikefjord Fault to be located within their claimed “Middle Plate”, and interpreted the fault as a *detachment* fault, i.e. the authors apparently ignored its steep orientation and brittle character.The issue of strike-slip faulting was not discussed. Johnston et al. (2007b) grouped the Eikefjord fault, and the Standal fault, as “high-angle normal and strike-slip faults”. From fault-slip analyses of meter-scale fault planes near the two faults, the authors found indications that initial E-W stretching and vertical thinning had been strongly overprinted by E-W stretching and N-S shortening. The authors suggested this to be consistent with “early E-W to SE-NW stretching followed by late sinistral shear”.

The southern boundary of the Eikefjord Group is defined by the Sunnar Fault (described below in Sect. 3.7) which represents the contact towards the Lower Palaeozoic rocks of the Høydalsfjorden Complex. The latter complex consists of metasediments and basic intrusions that have been interpreted as part of the primary cover sequence to the Solund–Stavfjorden Ophiolite Complex (Furnes et al. 1990), but which may also be equivalent to the Sunnfjord Melange or to the Kalvåg Melange (this work).

Regional correlation

The rocks of the Eikefjord area are part of a belt of "Anorthosite-Jotun Kindred" rocks (Goldschmidt 1916). This belt constitutes the areas ranging from the Gloppen district, located to the east of the Hornelen Devonian Massif (**Fig. 2.5**), and southwestwards via Eikefjord, to the Florø penninsula (**Figs. 2.3** and **2.4**) (Bryhni et al. 1981, Bryhni 1989). The western part of this belt can be observed on the 1:250.000 geological map sheet Måløy (Kildal 1970), and the eastern part may be studied on the 1:250.000 geological map sheet Årdal (Lutro & Tveten 1996). The meta-anorthositic rocks of the Eikefjord Group are also located in a narrow belt on the southern side of the Standal Fault just to the south of the Håsteinen Devonian Massif (**Fig. 2.5**) (Bryhni & Lutro 1991a, 1991b, 2000a, 2000b). The orthogneisses of the Eikefjord Group and the meta-supracrustal rocks of the Lykkjebø Group constitute the Fjordane complex (Bryhni et al. 1981; Bryhni 1989). The Eikefjord Group has been assigned to the Middle Allochthon (Bryhni & Sturt 1985; Andersen & Jamtveit 1990; Johnston et al. 2007b), whereas the Lykkjebø Group has been assigned to the (i) Lower Allochthon (Bryhni & Sturt 1985), and the (ii) Middle Allochthon (Andersen & Jamtveit 1990; Johnston et al. 2007b). Furnes et al. (1990) included the rocks of the Lykkjebø Group in their Høyvik Group, and the rocks of the

Eikefjord Group were included in their Dalsfjord Suite. (See Sect. 2.3 and Sect. 2.5). Recent accounts of the rocks in the Eikefjord–Gloppen area have been given by Young et al. (2007) and Johnston et al. (2007b).

Age

The gneisses in the Eikefjord Group are generally considered to be derived from Precambrian protoliths. A radiometric dating (Rb/Sr whole rock) has been performed on various meta-anorthosites and associated rocks in the Eikefjord and Gloppen area (**Fig. 2.5**), yielding an age of **1511 +/- 64 Ma** (Abdel-Monem & Bryhni 1978, see Ch. 2). In addition to this dating, it is generally accepted that lithological correlation with the same characteristic rock types of the Jotun Nappe Complex and the Dalsfjorden Nappe is valid (Bryhni and Sturt 1985). The two latter areas yield similar Precambrian ages (e.g. Schärer 1980; Corfu & Andersen 2002; Lundmark et al. 2007), which further supports a Precambrian age for the Eikefjord Group protoliths.

As mentioned above, Berry et al. (1993; 1995) and Andersen (1998) performed $^{40}Ar/^{39}Ar$ dating on white mica from both sides of the Eikefjord fault, obtaining an age of **~415 Ma** on the northern side, and **~ 400 Ma** on the southern side. However, the $^{40}Ar/^{39}Ar$ muscovite dating at **403 Ma** (Chauvet & Dallneyer 1992) just to the north of the fault, and the fact that the **~415 Ma** age is anomalous compared to numerous other datings in the Hornelen–Håsteinen area, indicates that the ages are the same on both sides of the fault.

Johnston et al (2007b) performed Sm-Nd dating on garnet obtained from garnet-muscovite scists belonging to the Lykkjebø Group. Samples were taken from three localities, and both cores and rims of the garnets were dated. A sample from near Svarthumle east of Eikefjord yielded a core age of **425.1 +/- 1.6 Ma**, and a rim age of **415.0 +/- 2.3 Ma**. A sample from Standal gave a core age of **422.3 +/- 1.6 Ma**, and a rim age of **407.6 +/- 1.3 Ma**. A sample from Austre Hydalen east of Hyen yielded only a rim age, at **414.1 +/- 1.6 Ma**. The core ages of **425–422 Ma** was interpreted as dating the initiation of peak *upper amphibolite* facies metamorphism in the rocks, which from thermobarometric analyses were estimated to have corresponded to **13–18 kbar** (**45–60** km depth). The rim ages of **415–407 Ma** was interpreteded to mark the end of these peak amphibolite facies conditions. Subsequent to these peak metamorphic conditions, the Nordfjord– Sogn Detachment Zone (NSDZ) was believed to have developed at **8–12 kbar** (**30–40 km** depth) conditions, i.e. *lower amphibolite–greenschist facies*, a development occuring after **~410 Ma** (mean rim age). Assuming that most of the movements on the NSDZ occurred during lower amphibolite faces, the published $^{40}Ar/^{39}Ar$ ages of the area, noted to be in the interval **402–396 Ma** (mean **400 Ma**), was taken to represent the youngest age of movements on the NSDZ. Accordingly the authors suggested that the movements on the NSDZ occurred in the time interval **410–400 Ma**.

3.3 ROCK DESCRIPTIONS

3.3.1 GENERAL

The rocks in the road sections in the studied part of the Eikefjord Group can be separated into two categories; (1) the completely dominating orthogneisses, and (2) subordinate meta-sedimentary micaschists. According to Bryhni et al. (1981) and Bryhni (1989), such micaschists actually belong to the Lykkjebø Group, but generally it is not uncommon that layers of these meta-sediments locally occur within the orthogneisses of the Eikefjord Group. For simplicity, however, all rocks in the area have here been placed under the heading "Eikefjord Group", since the orthogneisses are so dominating in the area. The first category, i.e. the orthogneisses, contains meta-anorthosites, biotite-epidote gneisses, intermediate to felsic augen gneisses, amphibolites and meta-gabbros, and dominates the area. In addition, ultramafics are present in the same rock association in other areas (Bryhni et al. 1981). The second category consists of two E-W trending belts of meta-sedimentary micaschists. The belts are **100–250 m** wide, and are separated by **about 250 m** in the study area. The micaschists can best be studied in the road section Sunnarviken–Osa, but will be only briefly mentioned here. For a comprehensive petrographic description of all rock types in the rest of the Eikefjord Valley, see Bryhni (1958). The rocks are not differentiated on the main map (**Plate 1**).

Due to the extensive mylonitisation, intrusive relations and other features showing relative ages between the rocks have been obliterated, and these subjects are therefore not further discussed in the present work. The narrow belt of Eikefjord Group meta-anorthosites present along the southern side of the Standal Fault, just to the south of the Håsteinen Devonian Massif, are located outside the study area and will therefore not be treated in the present work. On the main map (**Plate 1**), these rocks are included in the designation "undifferentiated rocks to the south of the Standal Fault".

3.3.2 MYLONITIC META-ANORTHOSITES

The meta-anorthosites in the investigated area are extensively mylonitized and metamorphosed. On outcrop scale, interbanding with dark grey epidote-biotite gneisses, or possibly retrograded amphibolites, is very common (**Fig. 3.2** and **3.4**). **Fig. 3.2.** shows a typical field appearance of the rock. The banding varies from millimetres (**Fig. 3.4**) to metres (**Fig. 3.2**) in thickness and is due to transposition during extensive mylonitisation of the rocks during deformational phase D_2. (Premylonitic D_1 structures have not been observe in the study area, but Bryhni et al. 1981 reported such structures near the study area, see Sect. 3.4.1). The mylonite fabric is accordingly designated S_2 / L_2. Locally, larger bodies of meta-anorthosite without admixture of other rock types are present. Bryhni (1958) described anorthosite with increasing amount of dark minerals and gradual transition to gabbro in other parts of the Eikefjord–Gloppen area.

The main mineral in the meta-anorthosites is plagioclase. In the mylonitic portions, subgrains are albitic, as generally throughout the group (Bryhni 1958), and the larger porphyroclasts may range up to **An_{50}**. The total amount of plagioclase varies up to above **90 %** of the rock. Up to **15 %** white mica may be present. Zoisite and/or epidote can occationally constitute up to **25 %** of the rock. Chlorite is present, and can locally constitute up to **15–20 %** of the minerals. (**Fig. 3.5**).

In outcrop, the mylonite fabric is the most prominent feature of the rock, and ultramylonitic variants of the rock are common (**Fig. 3.6**). The colour is characteristically white or whitish on broken surface (**Fig. 3.2**), and more grey-white on weathered surface. In outcrop, the meta-anorthosite is often completely white with no trace of dark minerals. Locally, shady discontinuous grey-coloured ribbons less than **1 mm** thick can be present in the otherwise white rock (**Fig. 3.6**). The meta-anorthosite is generally very monotonous with relatively little variations in mineralogy, colour and textural features. The grain size, however, does vary, and the rock shows some variations in strain intensity. Under the microscope, the ultramylonitic texture is seen to be defined by elongated plagioclase subgrains (**Fig. 3.7**). Less mylonitized versions of the rock are also common.

The protolith to the meta-anorthosites cannot be observed in the investigated road profiles, but Bryhni (1958) described primary violet-coloured labradorite (**An_{50-60}**) in coarse grained unmylonitized anorthosites at a small pond to the NE of the Knapstad area (**2 km** east of the village of Eikefjord, UTM ca. 145 328). This indicates that lenses of anorthosite locally may have escaped the extensive mylonitisation otherwise so characteristic for the Eikefjord Group.

3.3.3 MYLONITIC GREY GNEISSES

The grey gneisses constitute large parts of the studied area, and can conveniently be studied at several localities along the road sections.

At outcrop scale, the grey gneiss is either closely interbanded with meta-anorthosite (**Fig. 3.2** and **3.4**) or present as larger bodies (**Fig. 3.8**). The colour of the rock-type which is here termed grey gneiss varies considerably, ranging from dark grey-black to light grey. The colour becomes lighter grey with increasing amount of feldspar in the matrix. Upper part of **Fig. 3.8** shows the typical field appearance of the rock.

Bryhni et al. (1981) used the term biotite-epidote-gneiss for these rocks, and this name is considered appropriate also by the present author. The main minerals in the rock are plagioclase, K-feldspar, epidote, biotite, white mica and chlorite (**Fig. 3.9**). Plagioclase subgrains in the mylonitic matrix generally have an albitic composition, but the porphyroclasts may range up to above **An_{30}** (Bryhni 1958). Assessory minerals are actinolite, garnet, sphene, allanite and ore. The quartz content of the rock is negligible. Feldspars with exolution lamellae are common. These feldspars are usually porphyroclasts, and the grain size-reducing processes caused extensive subgrain development from these porphyroclasts (**Fig. 3.10**). Viewed through a polarizing microscope, the porphyroclasts have developed mantle-core structures and show long tails of feldspar subgrains that recrystallized during the mylonitisation (**Fig. 3.10**, **3.11** and **3.12**).

The rock is generally mylonitic, although the strong fabric can occasionally be difficult to discern at the outcrop due to lack of contrasting lithologies in the rock and, hence, vague banding. In such instances the rock can appear to be massive and structureless, and trains of feldspar augens will often be the only feature defining a visually detectable fabric. The rock can be observed in different stages of mylonitisation. Protomylonitic versions are present locally (**Fig. 3.13**), whereas mylonitic variants are more common (**Fig. 3.14**), as is the ultramylonitic version (**Fig. 3.15**). The grain size in the matrix is usually less than **0.3 mm**.

The grey gneisses are characterized by variable amounts of white feldspar augen which are either arranged in trains or scattered evenly throughout the rock. The clasts are either eye-shaped or more circular. The augen are interpreted as porphyroclasts which resisted the grain size-reducing processes during mylonitisation (cf. **Fig. 3.10**). The surrounding matrix-fabric bends around the clasts (**Fig. 3.12**). Occationally, the augen show signs of brittle cracking (**Fig. 3.16**). Eye-shaped clasts usually have their longest dimension oriented in the mylonite foliation (**Fig. 3.16**), but they can also be oriented with longest dimension orthogonal to it (**Fig. 3.15**). The porphyroclasts range in size from millimetres and up to several centimetres. The largest porphyroclast recorded measured **7 x 4 cm**, indicating a pegmatitic protolith. The augen in these gneisses were generally determined to consist of both K-feldspar and plagioclase (Bryhni 1958).

The rock is occasionally banded (**Figs. 3.8**, **3.14**, **3.15** and **3.16**), where the white bands are defined by higher content of feldspar. The thickness of the bands varies from a few millimetres to a few centimetres. The bands locally show a gradual transition to more isolated feldspar clasts, and can be both continuous and discontinuous on outcrop scale. The boundary between the bands and the surrounding grey gneiss can be both sharp and more transitional. Bands with thicknesses down to a few millimetres have the appearance of ribbons.

The protolith to this rock cannot be observed in the investigated profiles. Bryhni et al. (1981) reported that the grey gneisses generally have a syenitic to monzonitic composition throughout the Eikefjord Group.

Augen gneiss

An occurrence of augen gneiss is present in a limited exposure in the road section at the Vasset farm (UTM 1362 3050) (**Fig. 3.17**). The rock is generally distinctly different from the grey gneiss, as it is generally more felsic due to the characteristically high content of feldspar augens, but in the most strongly mylonitized parts, the augen gneiss may tend to approach the apparence of the grey gneiss. Since it can therefore not be excluded that the augen gneiss and the grey gneiss had protoliths with related bulk compositions, the augen gneiss is here presented together with the grey gneiss. The augen gneiss is not abundant in the studied area. In spite of this, it is included in the present description because of its ability to express spectacular L_2-stretching lineations (Sect. 3.4.2).

In outcrop, the rock usually has a fine-grained matrix with medium grey colour. In the mylonitic parts, the augen of white feldspar tend to be concentrated in numerous closely spaced bands. In less mylonitic parts, the augen are distributed more evenly, and may then comprise up to **40–50 %** of the rock. On foliation surfaces, the feldspar augen usually delineate a strong stretching lineation (**Fig. 3.17**). This lineation will be discussed in Sect. 3.4.2 (D_2-deformation). The augen in both the grey gneisses and the related augen gneiss were determined to generally consist of both plagioclase and K-feldspar (Bryhni 1958).

Rocks classified as augen gneiss are more common in the areas further to the NE, i.e. towards the Gloppen district, as shown for example on the Måløy map (Kildal 1970). In those districts the rock is easily separated from the grey gneiss.

3.3.4 AMPHIBOLITES

In the N-S trending road section located in the middle of the U-shaped road bend in Sunnarviken (**Plate 1**) (UTM 1215 3185), a lens of retrograded amphibolite is present. The lens has escaped the mylonitisation process, as it is massive and unfoliated.The rock is thus worth a description, although it does not appear to be widely represented in the study area. It should, however, be noted that amphibolites can easily be mistaken for dark versions of the grey gneiss, and amphibolites may therefore be present elsewhere in the study area, especially as dark bands in the meta-anorthosites. Bryhni (1958) also described amphibolites from the study area.

The eye-shaped lens is about **5 m** long and **3 m** high when measured in the N-S profile. The length in the eastward direction is not known. However, since the N-S profile is orthogonal to the E-W trending stretching lineations in the surrounding mylonites, the long axis of the body is probably oriented E-W. The lense is surrounded by strongly mylonitized meta-anorthosite, and both the lens and the mylonite fabric is dipping about **30°** towards south. It is not known whether other lenses of similar composition are present to the east. On broken surface, the colour is dark grey, and the dark amphibole minerals can easily be seen on the outcrop.

The grain size in the lense is less than about **1 mm**. The rock is characteristically very monotonous with only minor textural and mineralogical variations. The texture of the amphibolite is unfoliated. (**Fig. 3.18**). The rock is dominated by hornblende which may constitute up to **60–70 %** of the rock. Other minerals are plagioclase, biotite, actinolite and chlorite. Ore is the most prominent accessory mineral. Hornblende and biotite are commonly partially retrograded to actinolite and chlorite (**Fig. 3.19**).

3.3.5 MYLONITIC MICASCHISTS

The micaschists treated here are meta-supracrustal rocks (Bryhni 1958; Bryhni et al. 1981) present within the orthogneisses of the Eikefjord Group. The schists form two E-W trending belts separated by meta-gabbro (see the geological maps of Kildal 1970, and Bryhni & Lutro 1991a, 1991b, 2000a, 2000b). At the Agledal area (UTM 185 295) situated about **4.5 km** to the east-southeast of Osa/north end of Vassetvatnet (or **3.5 km** SE of Storebru), Bryhni (1958) studied the eastward continuation of the same schist belts and concluded that the schists belonged lithologically to the Lykkjebø Group. Bryhni (1958) interpreted the micaschists to be part of his "Kvartsitt-avdelingen" ("the quartzite unit"), which was later called the Lykkjebø Group (Bryhni et al. 1981). The contact between the micaschists and the gneisses is characterized by the penetrative mylonitisation. Possible original sedimentary or intrusive relations are obliterated. The intermixing of the orthogneisses and the micaschists is related to the extensive shear movements that produced the mylonites (Sect. 3.4.2). In the study area, the micaschist can conveniently be studied in the road sections midway between Sunnarviken and Osa.

In outcrop, the schists have a medium grey colour on broken surface, and a lighter grey colour on weathered surface. The mesoscopic appearance of the rock is dominated by a penetrative development of shear bands (Sect. 3.4.2). A complex pattern of anastomosing shear bands gives the cleavage in the rock a "chaotic" appearance on the outcrop, occasionally giving the rock a phyllonitic character. On the centimetre scale, the shear bands isolate lenticle-shaped domains of less micaceous material. Locally this anastomosing pattern gradually gives way to zones with one dominant fine-grained ultramylonitic foliation. The thickness of such zones can be up to a few metres. Bryhni (1958) reported that this rock elsewhere in the Eikefjord and Lykkjebø Groups occasionally grades into a more gneissic variant.

Under the microscope it is seen that the main minerals in the micaschists are feldspar, white mica, chlorite, biotite, garnet, clinozoisite and quartz. The plagioclase typically has an albitic composition throughout the area (Bryhni 1958). Accessory minerals are calcite and ore. Garnet is only present locally, but can then be up to **4 mm** in diameter (this work). Commonly the garnet shows cracking and dismembering with associated formation of chlorite (**Fig. 3.20** and **3.21**). Shear-related mica fish are common in the schist (**Fig. 3.22**). The grain size varies, from large mica fish measuring up to **5 mm** from tip to tip, to small feldspar and quartz subgrains less than **0.1 mm** in diameter.

The characteristic mesoscopic mylonite features, as well as typical microscopic mylonite textures (Sect. 3.4.2), suggest that the micaschist must be classified as a mylonite, although it locally has the appearance of a phyllonite.

The “Eikefjord” micaschists treated here are different from the micaschists of the Høydalsfjorden Complex (Ch. 4), in that the former micashists are structurally completely dominated by the anastomosing shear band patterns indicating that strong shear movements have affected the rocks. The schists are also mineralogically different due to different metamorphic histories. Eikefjord schists show a *retrograde* development from a higher grade towards greenschist facies, as indicated by the presence of the garnets and biotite altering to chlorite (Sect. 3.5). In contrast, the schists in the Høydalsfjorden Complex are *prograde* to the upper part of lower greenschist facies (chlorite grade). This is further discussed in Sect. 4.4.

3.3.6 META-GABBROS

Meta-gabbroic rocks are present between the two belts of micaschists described above (Sect. 3.3.5), and extends eastward from the Sunnarvik area. The meta-gabbro is displayed on the maps of Kildal (1970) and Bryhni & Lutro (1991a,1991b, 2000a, 2000b). The rock is well exposed for example along the eastern part of the road Sunnarvik–Osa. The petrography of the meta-gabbro bodies in the Eikefjord Group are extensively described by Bryhni (1958), and a separate description will not be given here.

Some of Bryhni's (1958) observations are of general significance. Bryhni (1958) observed almost undeformed and unaltered gabbro at only one locality in the Eikefjord Valley, and this locality is situated at the small lake Lomtjern **200 m** to the north of the main road (route 5) about **3 km** east-southeast of the village of Eikefjord (UTM 153 323). This rock contains ortho- and clinopyroxene, amphibole and plagioclase of $\mathbf{An_{50}}$. Moderately deformed and retrograded meta-gabbro typically contains oligoclase ($\mathbf{An_{10\text{-}30}}$) and actinolite as early retrograde products. However, in the most common mylonitic types, the retrograde alteration process has usually continued even further, and the rock approaches an actinolite-chlorite-epidote-schist (Bryhni 1958).

3.4 STRUCTURAL GEOLOGY

3.4.1 INTRODUCTION

The present section deals with the structures and fabrics developed during the penetrative mylonitisation of the Eikefjord Group, and the (post-shear/late? syn-shear) folding of this fabric. The description is essentially based on observations along the road sections studied; Eikefjord–Sunnarvik, Sunnarvik–Osa and Osa–Vasset (the last one illustrated in **Appendix B: Road Log no. 4**), but the description also includes data from the area between the Sunnar Fault and the road Sunnarvik–Osa (**Plate 1**).

It should be noted that the numbering of deformation phases in the present thesis relate exclusively to each tectonostratigraphic unit and only reflects the structural history within this particular unit. Based on the present structural interpretation, this means that D_1-structures in one unit, e.g. the Eikefjord Group, do not correspond to D_1-structures in other units, e.g. the Høydalsfjorden Complex, and so on with D_2, etc.

The lithologies in the studied area have not been remapped by the present author (Sect 3.1). The area is covered by the 1:50.000 geological map sheets "Eikefjord" (Bryhni & Lutro 1991a, 2000a) and "Naustdal" (Bryhni & Lutro 1991b, 2000b), and the outcrop pattern of the main rock types can be found on these maps. In addition, some main features can be observed on the 1:250.000 map sheet "Måløy" (Kildal 1970). The meta-sediments present within the orthogneisses of the Eikefjord Group are an integral part of the mylonite package.

As indicated above, the deformation structures in the Eikefjord Group can be separated into two distinctly different styles: **i)** The first deformational style recorded is interpreted to have been produced by a non-coaxial, penetrative shear deformation. The deformation event that produced the structures is denoted D_2, and has resulted in the development of the mylonitic fabric in the studied area and, presumably, also in the rest of the Eikefjord–Gloppen region. The deformational planar fabric is denoted S_2. Intrafolial folds in the mylonitic fabric are termed F_2-folds, although they might have formed successively during different stages of the mylonite formation. Mineral stretching lineations are termed L_2. The mylonitisation process obliterates all older structures in the rocks in the studied profiles. However, since relicts of earlier tectono-metamorphic events are present in similar rocks outside the studied area (Bryhni et al. 1981), these earlier signatures are collectively assigned to a D_1-phase, which will not be given further treatment here. The D_2-mylonitisation is accompanied by various structural features which will be presented below in Sect. 3.4.2. **ii)** The second deformational style recorded is related to the folding of the D_2-fabric, which produced gentle to close F_3-folds. An axial planar S_3-fabric has not been observed. This deformation is termed D_3, and the D_3-structures are presented below in Sect. 3.4.3.

3.4.2 D_2-STRUCTURES

S_2-mylonitic foliation

The shear-related D_2-deformation has produced a mylonitic SL-tectonite. The mylonitic S_2-foliation itself is the most prominent D_2-structure. The foliation is commonly associated with a lithological banding, which is locally very intense (e.g. **Fig. 3.2** and **3.4**). The banding is mainly defined by distinctly different rock types. Mineralogical variations due to metamorphic segregations can, however, not be excluded, particularely in the grey gneisses.

The degree of mylonitisation of the rocks varies, and these variations are particularly well displayed in the less monotonous versions of the grey gneiss. Protomylonitic versions are present (**Fig. 3.13**), but mylonites (**Fig. 3.14**) and ultramylonites (**Fig. 3.15**) represent the most common strain state. Meta-anorthosite is commonly ultramylonitic (**Fig. 3.6**). Microscopically, the typical mylonite fabric in both meta-anorthosite and grey gneiss are dominated by feldspar porphyroclasts and matrix subgrains, with variable amounts of other minerals (**Fig. 3.7**, **3.10**, **3.11** and **3.12**).

Shear bands

Shear bands are commonly used as kinematic indicators in mylonites, and such structures were studied by numerous authors during the 1980s and 1990s (e.g. Berthe et al. 1979; Platt & Vissers 1980; Lister & Snoke 1984; Weijermars & Rondeel 1984; Platt 1984; Passchier & Trouw 1996). A comprehensive terminology was developed for various types of shear bands, and they were classified as being of «S-C type I or II», «extensional crenulation cleavage (ecc) type», «C'» , etc. Structures of apparently similar type could be given different names. In the present treatment, however, the structures are used as kinematic indicators, and since general agreement exists as to the interpretation of sense of shear from the structures, they will here just be termed «shear bands».

In large parts of the area, shear bands are not visible at outcrop scale, and the rock is here an SL tectonite characterised by the dominating mylonitic fabric (e.g. **Fig. 3.6**, **3.15**, and **3.16**). Shear bands may, however, be observed at the microscopic level, e.g. as bands confining mica fish (**Fig. 3.22**), as bands anastomosing throught the mylonite fabric (**Fig. 3.23**), or as "crenulation-resembling" fabrics crossing he dominating mylonite fabric at different angels (**Fig. 3.24**). On the mesoscopic scale, shear bands occasionally off-set lithological banding within the mylonite, forming foliation boudinage (**Fig. 3.25**) (see below). Shear bands are particularly abundant in the micaschists.

Mesoscopically, it is concluded that the shear bands show overall top-to-the-west sense of movement. The basis for this conclusion is a number of individual field observations which together leave the clear impression of this sense of movement (cf. **Fig. 3.25**). The microscopic shear bands in oriented samples (e.g. **Fig. 3.22**) also show westward directed movement. In some areas, the sense of shear is difficult to deduce. Shear bands with a clear top-to-the-east sense of movement have not been observed, but caution must be taken for shear bands present in steep F_3-fold limbs, as these bands may be erroneously interpreted as indicating top-to-the-east sense of shear.

Intrafolial F_2-folds

Intrafolial F_2-folds are present, although not very abundant. In the literature, such folds have been used for deduction of sense of shear (e.g. White et al. 1986), although causion must be exercised (e.g. Passchier & Trouw 1996). In the study area, folds in centimetre- and metre-scale have been observed. One example of a fold in centimetre-scale can be observed in the road profile Eikefjord–Sunnarvik (UTM 1189 3293, **ca. 800 m** to the south of the main road junction in the village of Eikefjord). The fold has one long and one much shorter limb, and the short limb is about **2 cm** (**Fig. 3.25**). The F_2-fold axis of the fold is oriented in the Y-direction of the overall finite D_2 strain ellipsoid, i.e. about N-S. The vergence of the fold is westerly, consistent with a top-to-the-west sense of shear.

An intrafolial fold in meter-scale is present in the same road profile (UTM 1189 3283, **ca 750 m** to the south of the main road junction in the Eikefjord village). The fold is isoclinal, recumbent, and the length of the upper short limb (as seen in the road section) is minimum **3 m**, measured from the hinge zone to the edge of the road cut (**Fig. 3.2**). Late shear movements have destroyed parts of the "upper limb" (**Fig. 3.2**). In the hinge zone, (**Fig. 3.26**), lithological banding may be traced from one limb to the other. It cannot be excluded that the fold is a sheath fold. However, although sheath folds are a common feature in shear regimes (Cobbold & Quinquis 1980), the necessary high degree of three dimensional control required to actually *prove* the presence of this fold type is often not present, and this is the situation also in the present case. For the fold to be a sheath fold, the "long axis" of the fold body would have to be oriented in the X-direction of the overall strain ellipsoid of the present shear system, i.e. about E-W. Due to the lack of three dimensional control at the exposure, the fold will not be used to deduce the direction of movement.

Bryhni (1962) reported that isoclinal recumbent folds were abundant also in the Lykkjebø Group in the Grøneheia area just to the north of the village of Eikefjord.

Asymmetrical foliation boudinage

Asymmetrical foliation boudinage and accompanying shear bands have been observed (**Fig. 3.25**). These structures give information on sense of shear (e.g. Platt & Vissers 1980; Hanmer 1986). The structures show top-to-the-west sense of movement.

Porphyroclasts

Porphyroclasts of feldspar are locally present in large quantities in the grey gneisses (**Fig. 3.8** and **3.16**). In the literature, the geometry of the porphyroclasts and their recrystallised mantles ("subgrain tails"), are frequently used as shear sense indicators in mylonites (e.g. Passchier & Simpson 1986; Van den Driessche & Brun 1987; Bjornerud 1989; Passchier & Trouw 1996). In the Eikefjord Group, porphyroclasts of the sigma-type clast system of Passchier & Simpson (1986) are commonly present (e.g. **Fig. 3.10**). When asymmetrical tail geometries are developed, sense of shear tends to be top-to-the-west. This is based on several tens of individual observations. Lack of well developed asymmetrical tail geometries are common, making deduction of vorticity difficult at times (e.g. **Fig. 3.11**, **3.12**). The feldspar porphyroclasts have occasionally been subjected to brittle cracking (**Fig. 3.16**). This may suggest that the last parts of the shear movements occurred under low grade metamorphic conditions, although composition contrasts and strain rate may have played a role.

L_2-stretching lineations

L_2-stretching lineations are present in all rock types, but can be particularly well studied in the augen gneiss at the Vasset farm (UTM 1362 3050) (**Appendix B: Road Log no. 4,** station 23) where the lineation is defined by elongated white feldspar porphyroclasts with associated tails (**Fig. 3.17**). The trend and plunge of the lineations at this particular locality is around **275/10**, which is representative for the stretching lineations in the area. The grey gneiss also occasionally displays strong "rodding-like" lineations (**Fig. 3.27**) which may be a result of rod-shaped "higher-concentrations" of more competent material (e.g. feldspar), and which are particularly prominent when occurring together with the "rodding"-parallel hinges of small F_3-folds. The structures may also occasionally resemble mullions.

Stretching lineations from the area have been plotted stereographically (**Fig. 3.28**). The average azimuth and plunge of the lineations are about **280/10**. The deviation from this average is small, and a separation into structural subareas has therefore not been necessary. The lineations in the thesis area are, when present, usually fairly strongly developed. The lineation directions are consistent with data reported by Bryhni (1958; and unpublished lineation map), and also by Chauvet & Seranne (1989) and Seranne et al. (1989); the latter two showing the lineations as arrows (without dip/trend values) drawn on regional map figures. Wilks and Cuthbert (1994) conducted an extensive study of lineations in the region, showing stereoplots of lineations recorded from **13 stations** covering an area extending from the town of Florø and east- and northeastwards to the Gloppen area. Their lineation trend corresponds well with the directions of the present thesis. Lineation measurements were also presented by Johnston et al (2007b) from the Eikefjord–Hyen area, again with orientations consistent with the previous ones. The orientation of the L_2-stretching lineations coincides with the F_3-fold axis (see discussion in Sect. 3.4.3 and Ch.6).

Rock architecture

Sheets of different rock types, with thicknesses from centimetres to tens and probably hundreds of metres, are present as slices in the mylonite package. No laterally predictable or logic three dimensional organisation of the slices appear to exist on a local scale. Recognition or correlation of single lithological units thus become impossible over even short distances. The area does not comprise tectonostratigraphic units that can be separated from each other by e.g. discrete mylonite zones (as in a thin-skin foreland framework). In contrast, the whole group is heterogeniously and penetratively mylonitized. This has created a very complex and constantly changing geometrical and spatial relationship between rock types in the mega shear zone. Due to the lack of lithological markers suitable for achieving controle on structural geometries, it has not been possible to establish neither a tectonostratigraphy within the Eikefjord Group, nor three dimensional control on the folding of the mylonite fabric (see Sect. 3.4.3).

Local structural thickness and minimum amount of displacement

Within the study area, the absence of laterally persistent structural markers and the absence of a clear envelope surface for the folded rock fabric, make it very difficult to estimate even the *local* structural thicknesses of the Eikefjord Group. In the road section Osa–Vasset (**Appendix B: Road Log no 4**), the more than **800 m** of relatively continuous exposures allow a structural thickness to be estimated to **about 550 m**. This is based on an average southward dip of **about 30°** for the mylonite foliation. (See also Sect. 3.4.3). Outside this profile, the lack of structural control or lack of consistently dipping foliations make estimation of structural thicknesses difficult. The estimate of **550 m** is thus a minimum value.

A rough estimate of the minimum amount of shear displacement necessary to produce the **550 m** of mylonites may be given. A conservative estimate would suggest that the production of mylonites requires a shear strain $\gamma \geq 10$ (Skjernaa 1980; Fossen & Rykkelid 1990; Swanson 1992), implying a minimum displacement of **5.5 km**. A less conservative shear strain estimate was given by Fossen & Rykkelid (1990), who suggested that mylonite-formation required $\gamma \geq 15$, implying a minimum displacement of **~ 8 km**. Calculations of displacement based on the total thickness of the NSDZ in the Eikefjord–Gloppen area are given below.

Summary on D_2-structures

The kinematic indicators that have been briefly described above show a consistent top-to-the-west sense of movement in the mylonite package. This sense of movement is based particularly on observations of a large number of shear bands and several tens of observations of porphyroclast-tail geometries. These data confirm observations of sense of shear reported by Chauvet & Seranne (1989) and Seranne et al. (1989) from the same area. The data also fit well with the extensive lineation data-base of Wilks & Cuthbert (1994), which was sampled in the *NSDZ* east and south of the Hornelen Devonian massif but north of the Eikefjord Fault. The thesis data are also consistent with the lineations reported by Johnston et al. (2007b) from the NSDZ of the Eikefjord–Hyen area.The observations of the present thesis are furthermore compatible with the data recorded and the conclusions drawn in various previous publications, e.g. (i) Andersen & Jamtveit (1990), who reported from the mylonites of the "Standalen Detachment", a segment of the *Nordfjord–Sogn Detachment Zone (NSDZ)* positioned along the southern margin of the Håsteinen Devonian Massif; (ii) Krabbendam & Dewey (1998), who made a detailed study of kinematic indicators in the *NSDZ* south of Håsteinen, at the westernmost exposures along the shore of Førdefjorden west of Standal**;** and (iii) several publications describing structures in the mylonites of the "Kvamshesten Detachment" (another segment of the *Nordfjord–Sogn Detachment zone*) below the Kvamshesten Devonian Massif. Structures showing top-to-the-*east* sense of shear have not been observed in the Eikefjord Group mylonites. Tail-geometries of porphyroclasts may locally be symmetrical, but the overall impression is nevertheless top-to-the-west sense of movement. No structural markers are present in the mylonites, and the three dimensional architecture of the lithologies cannot be established. A minimum structural thickness of the mylonites in the study area is estimated to **550 m**.

Regional extension of the S_2-mylonites

The Eikefjord Group appears to be mylonitic in the whole area between the Håsteinen and Hornelen Devonian massifs, i.e. between the Sunnar Fault and the Haukå Fault (**Fig. 2.6**). The author has made reconnaissance investigations along the road-sections from Grov (on the Florø peninsula) and northwards across the Nordalsfjorden bridge, continuing further eastwards to Grøndalen, and also westwards to Haukå (**Fig. 2.4**). Reconnaissance work has also been carried out in the rocks of the Lykkjebø Group along the road sections from Storebru (**Fig. 2.5**) and northwards along the lake of Endestadvatnet in the direction towards Hyen. In both of these areas the rocks are strongly mylonitic.

Several workers that have accompanied the thesis author on reconnaissance work in the area, or informed about their own reconnaicance or detailed work there, have supported the conclusion that the entire area is mylonitic and part of he NSDZ (pers. com. Haakon Fossen, Alan G. Milnes, Mike Norton). Recently, Johnston et al. (2007b) has also reached the same conclusions in their study of the Eikefjord–Hyen area.

Total structural thickness of the NSDZ between Hornelen and Håsteinen

The total structural thickness of the Eikefjord Group in the whole area between Eikefjord and Gloppen is difficult to estimate. This is due to the interlayering and interfolding of orthogneisses from the Eikefjord Group, with meta-supracrustals from the Lykkjebø Group. In addition, the envelope surface of the mylonite foliation is sub-horizontal, preventing a cross-sectional view of the zone. Put together, however, the two groups constitute the *NSDZ* mylonitic package in this part of Sunnfjord. Wilks & Cuthbert (1994) estimated the structural thickness of the mylonite zone in the Eikefjord–Gloppen area to **4–6 km**, but did not explain how this estimate was obtained.

Along the Standalen Fault to the south of the Håsteinen Devonian Massif, the NSDZ mylonite package was termed the”Standalen detachment” by Andersen & Jamtveit (1990), who estimated the structural thickness to be **2–3 km**. In this area, the whole shear zone generally has a consistent dip to the north, allowing a cross-sectional view of the zone, that makes it easier to estimate its structural thickness. It is believed that this thickness estimate of **2–3 km** is relevant also for the Eikefjord–Gloppen area, since the Standalen segment of the NSDZ continues into the Eikefjord–Gloppen area.

Amount of displacement on the NSDZ between Håsteinen and Hornelen

The amount of shear strain (γ) required to produce a mylonite has been estimated to $\boldsymbol{\gamma \geq 10}$ (Skjernaa 1980; Fossen & Rykkelid 1990; Swanson 1992). If the shear strain is combined with the **2–3 km** thickness of the shear zone in the Eikefjord–Gloppen area, the minimum amount of displacement on the zone may be calculated. A conservatively estimated shear strain of $\boldsymbol{\gamma \geq 10}$ (Skjernaa 1980; Fossen & Rykkelid 1990;

Swanson 1992) yields a minimum displacement of **20–30 km**. Applying a less conservative estimate of **$\gamma \geq 15$** (Fossen & Rykkelid 1990) would give a minimum displacement of **30–45 km**. Since the westward movement of the Devonian basins (and Upper Plate) started at an original position somewhere between the Jotun Nappe and their present coastal locations, the estimated displacement may be compared with measurements of the distance between the Hornelen Devonian Massif and the Jotun Nappe. These measurements may be taken from a point at the easternmost part of Hornelen, to obtain distances in slightly various directions towards the Jotun Nappe: the distance in the *SE-ward direction* is **80 km** (to Leikanger), which is the shortest distance between Hornelen and the Jotun Nappe; the distance in the *ESE-ward direction* is **100 km** (to Skjolden); and the distance in the *E-ward direction* is **140 km** (to Lom). Hence, if applying a **$\gamma \geq 15$** (displacement **$\geq$ 30– 45 km**) to the **80 km** direction to Leikanger, the mylonite-thickess of **2–3 km** estimated by Andersen & Jamtveit (1990) appears to suggest that the Hornelen basin could have originated minimum half way between its present position and the Jotun Nappe. However, in making such estimates of displacement magnitudes on shear zones, the validity of the estimate depends on the important condition that the shear zone is not a reactivated older shear zone, since such reactivation requires less shear strain than the formation of a new zone. As a matter of fact, Wilks & Cuthbert (1994) made such a suggestion, claiming that the shear zone in the Eikefjord–Gloppen area was in fact a top-to-the-west reactivation of the Scandian top-to-the-east thrust-related decollement zone. If this is the case, the above displacement estimate is not valid. However, other workers, including the present author, hold the shear zone in the Eikefjord–Gloppen area as part of the Nordfjord–Sogn Detachment Zone, which further to the south is believed to have cut through the Caledonian nappe stack (e.g. Milnes et al 1997). The nappe-cutting nature of the zone suggests that the zone does not follow Scandian nappe-bounding thrust mylonits, although a limited degree of reactivation of thrust mylonites cannot be excluded. Anyway, the very consistent and penetrative top-to-the-west fabric in the zone attests to a very high γ-value. In this perspective, the displacement estimate will be valid.

Significance of the S_2-mylonites and related structures

The S_2-mylonites and related structures in the Eikefjord Group is interpreted to have been formed as part of the mylonitic *Nordfjord–Sogn Detachment Zone*. On the basis of the reconnaissance work referred above, it is suggested that the whole area between the Håsteinen and Hornelen Devonian Massifs (defined by the Eikefjord and Lykkjebø Groups), as well as the eastern part of the Florø penninsula (**Fig. 2.3**), are part of the Nordfjord–Sogn Detachment Zone. Also the whole area a bit further east, between the Standal Fault and the Haukå Fault (**Fig. 2.3** and **2.6**), is probably part of the detachment zone, as briefly suggested by Andersen & Jamtveit (1990) and more thoroughly substansiated by Wilks & Cuthbert (1994) and Johnson et al. (2007b).

On the profiles (Plate 2), the detachment zone is shown without internal structures, since the structural geometry cannot be inferred due to lack of lithological markers. The mylonite zone was formed between the western Upper Plate and the eastern Lower Plate during the Devonian extensional movements. These movements eventually led to the deposition of the Devonian sediments on the Upper Plate (see review of the "detachment" model in Sect. 2.8). The detachment zone is generally **2–3 km thick**, (probably not **4–6 km** as suggested by Wilks & Cuthbert 1994 for the Eikefjord–Gloppen area), and is the result of extreme shear movements. In the literature, the minimum displacement on the zone has varied between **40** and **60 km** (for example **50–60 km** by Andersen & Jamtveit 1990; **40–50 km** by Wilks & Cuthbert 1994 for the Eikefjord–Gloppen area). In the study area , the intermixing of "sheets" of meta-sedimentary micaschists of the Lykkjebø Group with orthogneisses of the Eikefjord Group is related to the shearing within the zone.

Wilks & Cuthbert (1994) presented stereo plots of a large number of streching lineations and mylonite foliation poles, demonstrating that the lineations were essentially horizontal and the foliation poles essentially positioned along N-S girdles. (The foliation pole girdles reflected the folding of the mylonite fabric about subhorisontal E-W trending fold axes). The same results were obtained by Johnston et al. (2007b).This orientation of lineations and mylonite foliations implies that the envelope surface of the detachment zone in this large area, is oriented sub-horizontally, and the topographical relief in the area displays different levels of the zone. Also the **2–3 km** thick zone of mylonites located just to the south of the Standal Fault is interpreted as part of the Nordfjord–Sogn Detachment Zone, and Andersen & Jamtveit (1990) called this segment the "Standal Detach-ment". The nature of the Nordfjord–Sogn Detachment Zone is further discussed in Ch. 6.

Does a "Middle Plate" exist?

Andersen & Jamtveit (1990) suggested that the whole Eikefjord–Gloppen area constitutes a separate plate — termed the “Middle Plate” — situated between the Upper Plate (Hornelen Devonian sediments and its substrate) and the Lower Plate (WGR). The consept of a “Middle Plate” has later been applied in several other publications (e.g. Dewey et al. 1993, there called “Middle Tectonic Unit”; Andersen et al. 1994; Berry et al. 1995; Andersen 1998; Krabbendam & Dewey 1998). In the “Middle Plate” model, the Nordfjord–Sogn Detachment Zone is thought to be located *underneath* the “Middle Plate”. In map-picture, this would imply that the NSDZ would crop out along the present eastern "margin" of the Eikefjord–Lykkjebø Group rocks. The upper contact of the “Middle Plate” is thought to be the so-called Hornelen Detachment, being defined as the brittle fault that is presently separating the eastern part of the Hornelen Devonian Massif from the subjacent mylonites. The Eikefjord Fault was by Andersen & Jamtveit (1990) claimed to be a separate major shear zone within their “Middle Plate”, and subsequently, Berry et al. (1995) suggested that the Eikefjord Fault was a “major detachment fault”. Both these works apparently ignored the fact that the structure is a brittle steep fault.

As explained above, the Eikefjord- and Lykkjebø rocks between Eikefjord and Gloppen are penetratively mylonitised, containing abundant top-to-the-west shear sense indicators. Therefore, the whole area is clearly part of a mega shear zone, which has also been convincingly demonstrated by Wilks & Cuthbert (1994) and recently confirmed by Johnston et al. (2007b). This shows that there is no reasons to maintain the concept of a “Middle Plate”. Accordingly, the conclusion drawn in the present thesis, is that no “Middle Plate” appears to exist. In other words, the whole area between the Håsteinen and Hornelen area can be incorporated in the NSDZ.

The NSDZ between Håsteinen and Hornelen: a Mode I or Mode II zone?

Wilks & Cuthbert (1994) suggested that the Nordfjord–Sogn Detachment Zone in the Håsteinen–Hornelen area corresponds to the original Caledonian basal thrust decollement zone, which in the area was seen as having subsequently experienced reversed movements, changing it to an extensional detachment zone. This type of detachment zone would correspond to the **Mode I** zone of Fossen (1992). However, most authors (e.g. Fossen 1992; Milnes et al. 1997) have claimed that the mylonite zone in the area corresponds to a **Mode II** detachment zone, i.e. a zone cutting through the Caledonian nappe pile and into the basement. This model is supported by several lines of evidence: **(i)** $^{40}Ar/^{39}Ar$ ages are **~ 400 Ma** in the detachment zone and Lower Plate (Andersen 1998), and **~ 450 Ma** in the Caledonian rocks of the Upper Plate (e.g. Andersen et al. 1998). If the NSDZ did not cut the nappe pile, the exhumation of the WGR (Lower Plate) would have been due to erosional unroofing only, with no contribution from tectonic unroofing. Erosional unroofing is a slower process, that would have given younger $^{40}Ar/^{39}Ar$ ages than the **400 Ma** recorded in the NSDZ and the Lower Plate. **(ii)** The mylonite package of the NSDZ is not a ramp in a Mode I zone, because the Caledonian nappe rocks involved in the NSDZ is generally highly thinned, implying that the shear zone is cutting up section across the nappe pile.

The interpretation of Wilks & Cuthbert (1994), that the mylonites were a Mode I zone, would have consequences for the southward continuation of the NSDZ in the Kvamshesten and Solund districts, where the Mode II character of the zone seems clear. However, Wilks & Cuthbert (op.cit.) did not discuss the consequences of their Mode I interpretation, for the situation further south.

3.4.3 D_3-STRUCTURES

General

The D_3-structures are represented by F_3-folds that fold the D_2-mylonitic fabric. Orientations of fold axes, axial planes (abbreviated "FAs" and "APs") and also folded S_2-foliations (i.e. F_3-limbs and hinge zones) have been recorded in the road profiles investigated; particularly in the Eikefjord–Sunnarvik profile and the Osa–Vasset profile (the last one is shown in **Appendix B: Road Log no. 4**), but also from the area located between the Sunnarvik–Osa road and the Sunnar Fault. The structural elements are plotted in stereograms (**Fig. 3.29**, **3.30** and **3.31**). The F_3-folds have not developed an axial plane cleavage, and the lack of such cleavage indicate that the D_3-deformation did not lead to significant recrystallisation of the S_2-mineral fabric.

In the road sections, the interlimb angles of the F_3-folds are gentle to close. The wavelengths varies from **a few centimetres** to **20 m** with average of about **6 m**. The amplitude varies from **a few centimetres** to about **3 m**. Possible larger-scale folds cannot be observed, but may be inferred (see below). The F_3-folds are always observed to fold the S_2-fabric, and the mylonite-producing D_2-shear movements have not been observed to cut the folds.

The Eikefjord–Sunnarvik profile

FAs and poles to APs in the road profile between Eikefjord and Sunnarviken are plotted in **Fig. 3.29a**. The average FA has an azimuth/dip of **WNW/10°**, and the APs appear to dip mainly towards S, although APs with opposite dip are also present. The poles to the APs define a great circle around the FA, i.e; a π-axis would correspond to the measured FAs. (The spread of AP poles is discussed in a separate section below). Poles to foliation surfaces, i.e. to the limbs and hinge areas of the F_3-folds, are plotted in **Fig. 3.29b.** The poles define a great circle, also around the FAs.

The area from the road Sunnarvik–Osa and southwards

FAs and poles to APs in the area between the Sunnar Fault and the road Sunnarvik–Osa have been plotted in **Fig. 3.30a**. Although the amount of data is limited, the FAs have a gentle plunge towards the west, and the axial planes appear to lie on the same great circle as for the Eikefjord–Sunnarvik profile. (The spread of AP poles is discussed in a separate section below). Reconnaissance in the area verifies that the FA orientation is consistent for the area. Foliation surfaces have mainly been aquired from a belt along the Sunnar Fault. The poles to these surfaces have been plotted in **Fig. 3.30b**, and the foliation has a gentle to moderately southward dip. The poles appear to lie on the same great circle as the Eikefjord–Sunnarvik foliation poles.

The Osa–Vasset profile

FAs and poles to APs in the road profile Osa–Vasset have been plotted in **Fig. 3.31a** (see also **Appendix B: Road Log no. 4**). The Osa–Vasset FAs have essentially the same orientation as both the Eikefjord–Sunnarvik FAs and the FAs in the area between Sunnarvik and Osa–Vasset. The Osa–Vasset APs, however, dip mainly to the north, although the poles show a slightly greater spread than on the two previous plots. The spread is interpreted as a primary feature caused by competence contrasts between different rock types, (The spread of AP poles is further discussed in a separate section below). If more data on AP poles were available, the APs would probably once again define a great circle which would have a π-axis corresponding to the measured FAs. This is likely, since also southward-dipping axial planes are obviously present in the Osa–Vasset profile, as seen for example at the **720 m** point of the profile (**Appendix B: Road Log no. 4**), although the position of the folds high up in the vertical road-section made recording of orientation impossible at this particular locality. The fact that the north-dipping axial planes in the profile Osa–Vasset appear to dip the opposite way of the majority of those in the Eikefjord–Sunnarvik profile, is probably the result of local strain variations. Poles to foliation surfaces defining the F_3-folds are shown in **Fig. 3.31b**. The poles again define a great circle around the gently westward dipping FAs, but with an overweight of foliations dipping to the south.

Vergence of F_3-folds and dip of envelope surfaces in the Osa–Vasset profile

The road profile Osa–Vasset has fairly continuous exposures over a distance of more than **800 m**, and a road log of the main structural features has been obtained (**Appendix B: Road log no. 4**). The log summarises data on general foliation orientation, fold morphology, and fold vergence. The road section shows several interesting features, and the fold styles in the log will first be closer described to draw attention to these F_3-folds.

In addition, it is briefly discussed why the vergence of these F_3-folds does not yield information on the existence of a possible major anticline between the Håsteinen and Hornelen Devonian Massifs. The background for a brief discussion on the presence of such an anticline is as follows: both the Håsteinen and Hornelen Devonian Massifs are clearly *synclines*. The "Solundian Orogeny" model, suggesting that a "Solundian Orogeny" (see Sect. 2.9) is responsible for the deformation of the Devonian rocks into synclines, states that a corresponding large-scale *anticline* is situated in the area between the two massifs. Such anticlines have also been proposed in other models, e.g. the model of Krabbendam & Dewey (1998), where sinistral movements along the MTFZ, combined with transtension in western Norway, are believed to produce N-S shortening in these areas. With a more or less continuous **800 m** profile, the vergence of F_3-folds could be of particular interest in an attemt to test this model. The problems related to geometric control of the fold pattern, the fold vergences and the possible presence of an anticline to the north will be discussed with reference mainly to the limited study area.

Two fold styles are present in the profile (**Appendix B: Road log no. 4**). The two fold styles (or *fold sets*) should be considered as end members, as transitional styles occur. *The first set of folds*, which dominates the profile, consists of large, asymmetrical, gentle folds with one long limb and one comparatively short limb, i.e. folds with a clear vergence (Bell 1981). The folds have wavelengths of more than **20 m** and amplitudes in the range of **3 m**, but folds with smaller wavelengths and amplitudes are also present. In the profile, mainly long limbs to these folds are portrayed, but where the hinge zone is seen, a wavy fold pattern appears in the profile. In the log interval **150–230 m**, a ‘complete’ fold (‘long limb – short limb – long limb’) with its hinge zone is exposed. The short limb is positioned in the interval **185–200 m**. The fold might give the impression of having no vergence because of the limited horizontal and verical exposure. However, by extrapolating limbs which are present to the north and south of the fold knee, it evidently is a *Z-fold* (viewed westwards) climbing towards the N. The same vergence is also present for most of the other folds of this type in the profile. *The other set of folds* consists of generally smaller, asymmetrical, close folds with wavelengths and amplitudes in the range of **2 m**. They are likely to be parasitic folds on the larger first set folds. Since this fold type is also asymmetrical, fold vergence is again present (Bell 1981). The fold style is well represented in the profile in the intervall **250–300 m**. These folds have an *S-shape* (viewed westwards), i.e. opposite to the first set which has Z-shapes. The S-folds appear to occur on the southward-dipping long limb of the large Z-fold that was discussed above. However, parasitic folds in this position should — according to conventional theory (Bell 1981) — have been Z-folds. It should be noted that the apparant *southward* dip of *axial planes* of the folds in the **250–300 m** interval of the profile is merely an effect of oblique intersection of the FAs that trend **090–100°**

and the road-cut that strikes **ca 045°**. In reality, the APs in the profile <u>do</u> have *N- or NNW-ward dips*, as shown in **Fig. 3.31a.**

Both fold styles are considered to be F_3-folds, and this shows the variety of F_3-fold styles represented. The FAs plotted in **Fig. 3.31a** come from folds belonging to both styles in the road section, and the FAs are seen to have the same orientations. Both fold styles also have APs dipping mainly to the north (**Fig. 3.31a**). The two fold styles therefore cannot be separated on the bases of orientation of structural elements, and this suggests that they are both F_3-folds.

The profile also yields information on the overall orientation of the mylonite foliation. The mylonite fabric is mainly dipping **20–40°** to the south, i.e. directed underneath the Håsteinen Devonian Massif. Considered separately, this dip direction may be taken to indicate an anticline to the north, and a syncline to the south below the Håsteinen Devonian Massif. However, when the profile is studied in detail, a greater complexity is revealed: As mentioned above, folds with opposite senses of vergence are present in the profile. The *Z-folds* (viewed westwards), with one *long* limb (which dips towards the south) and one *short* limb, are stepping *up* towards the *north*, while the *S-folds*, that appear to lie on the south-dipping limb of the large Z-fold, are stepping *down* towards the *south*. In the profile, way-up criteria do not exist, and neither does a laterally persistent rock sequence/association or structural marker lithology/horizon. Instead, the different lithologies are sheets, bands or lenses transposed into parallel orientation during the shearing. The lithologies taper out and have limited extent within the mylonite fabric. Hence, no correlatable and recognizable structural or lihological "levels" exists over distances larger than a few tens of metres, that is, until they dissapear out of the exposure. This means that it is impossible to observe whether a certain structural level or lithology is repeated throughout the profile. Envelope surfaces of the folds, and the general foliation dip essentially towards the south, and this means that possible higher order folds must be overturned towards the north with limbs dipping southwards.

<u>Discussion</u>**:** The present discussion illustrates the difficulties arising when attempting to use the vergence of F_3-folds and the dip of envelope surfaces to locate possible higher order folds in the profile:

The S- and Z-folds could theoretically be claimed to be parasites on opposite limbs of folds of higher order. Localisation of limbs of such higher order folds would, however, not be trivial, since marker lithologies to link up the fold limbs cannot be recognised in the profile. As mentioned above, a more straight forward interpretation of the profile would be that the “parasitic” S-folds appear to be positioned on the long limb of a larger-scale Z-fold. Such a fold combination is contradictory to conventions on fold vergence (Bell 1981), stating that “parasitic” folds have the same vergence as folds of higher order, i.e. smaller-scale S-folds should be parastits on larger-scale S-folds. Pfaff & Johnson (1989, and references therein), have shown that folds of opposite vergence, i.e. S- and Z-folds, may develop within the *same* limb of a higher-order fold, when layer-parallel shear is operating. However, since their model requires one of the fold sets to be monoclinal kink folds, and both fold sets to be of the same order, the model is not applicable to the Osa–Vasset folds.

To solve the problem of defining and separating limbs of possible higher order folds, the presence of lithological or other recognisable structural markers are crusial. But, as these are abscent in the present area, the problem cannot be easily solved. Even if possible higher order folds in the profile *could* be located, and the S- and Z-parasites were assigned to *opposite* limbs, this would not be sufficient to indicate an anticline towards the north. The reason for this is that a lithological/structural marker would be needed to document that the folds were *indeed stepping upwards* towards the north, with an envelope surface *indeed rising* towards the north. Without such a marker, the folds could e.g. have a horizontal envelope surface, and thus be parasitic folds on e.g. a recumbent higher order fold, and no anticline would then be present to the north.

***Conclusions*:** It has been shown that the absence of a marker horizon, combined with folds with opposite vergences, means that the presence of a southward-dipping foliation/envelope surface in the present profile is *not* sufficient to say that an anticline, corresponding to the Osstrupen Syncline, is present to the north. The lack of structural markers makes it methodologically problematic to test whether such an anticline is present. The Eikefjord–Sunnarvik profile (which runs roughly parallel to the Osa–Vasst profile dicussed here), cannot either contribute to solve the problem, as this profile have relatively few and discontinuous exposures, and suffer from the same lack of structural markers. In this profile, the general foliation is dipping both towards the south and north, and the folds alternate between having S- and Z-shapes along both north and south dipping envelope surfaces. Hence, conclusions on overall vergence cannot be drawn. No conclusive evidence is therefore present in the study area for an anticline towards the north.

In the Grøneheia area (**Fig. 2.5**) to the north of the village of Eikefjord, Bryhni (1962) and Bryhni & Lutro (1991a, 2000a) have documented the presence of an E-W oriented gentle *synform* spanning a width of at least **5 km** of the total N-S distance of **10 km** between the Håsteinen and Hornelen Devonian Massifs. This shows that the region between the massifs is not one simple antiform (-cline). It thus remains to be documented that anticlines corresponding to the Devonian synclines are really present between the Devonian massifs of Håsteinen and Hornelen, as assumed by the "Solundian Orogeny"-model (Sect. 2.9), and various other models. The "detachment model" (Sect. 2.8), however, does not have to imply such an anticline. In this model, the areas between the Devonian massifs may be considered to be structural ramps/highs/ in an irregular wavy/corrugated detachment zone, and the opposite dips of the foliations along the opposing sides of such highs, which make the highs resemble "anticlines", is assumed to reflect these primary irregularities in the shape of the detachment zone. But, regardless of whether a large-scale anticline is present between the Håsteinen and Hornelen Devonian Massifs or not, the smaller-scale F_3-folds in the Eikefjord Group must nevertheless have been formed due to some sort of N-S contractional forces. Large-scale sinistral wrench tectonics to the NW of present western Norway, at the end of the extensional phase, may have caused this slight crustal shortening (Sect. 2.8). This is further discussed in Ch. 6.

The spread of AP poles of F_3-folds

The AP poles to F_3-folds show a great spread (**Fig. 3.29a, 3.30a, 3.31a**), which is particularly well illustrated in the Eikefjord–Sunnarvik profile (**Fig. 3.29a**) where the poles are distributed along a large part of the great circle. The fanning of AP poles could theoretically arise from at least five processes, being: **(i)** primary fanning of APs due to lithological or rheological heterogeneities, **(ii)** N-S shearing during folding, **(iii)** local vertical flattening, **(iv)** σ_1- and σ_2-pertubations during folding (vertical σ_1 and horizontal σ_2 give horizontal AP; vertical σ_2 and horizontal σ_1 give vertical AP); **(v)** two different folding events, implying presence of different stress systems with e.g. similar σ_1 orientation and sligtly different σ_2 and σ_3 orientations. From the data, it appears that alternative (i) “primary fanning of APs due to lithological or rheological heterogeneities”, is a likely explanation. This may be appreciated by comparing APs in the two roughly N-S oriented profiles: above, it was noted that APs in the Eikefjord–Sunnarvik profile mainly dip ***S-wards***, whilst and Osa–Vasset profile have APs mainly dipping ***N-wards***. The two profiles are located straight opposite to each other, i.e. with FAs and lineations trending from one profile into the other. The distance between the profiles is only **2–3 km**. This change in foliation-dip suggets that the AP orientations display unsystematic variations even within relatively small areas, indicating that the fanning is a primary feature. Contributions from the four other alternatives cannot be excluded, but are less likely.

Summary of D_3-structures

The F_3-folds are oriented with W to WNW trends and shallow plunges. The poles to APs define great circles around the FAs. Poles to S_2-foliations, i.e. to limbs and hinge zones of F_3-folds, also define great circles around the FAs, and the majority of the surfaces recorded have a dip to the south. Both S- and Z-folds (viewing west) are present in the Osa–Vasset profile, but absence of structural markers makes it impossible to use fold vergence and the dip of envelope surfaces to deduce the positions of possible larger-scale folds. No evidence is therefore present for an anticline to the north corresponding to the Osstrupen Syncline in the south.

Timing of the F_3-folding

In Sect. 3.4.2 (on D_2-structures), the S_2-mylonites and the related L_2-lineations were interpreted to be part of the Nordfjord–Sogn Detachment Zone, which developed as a result of Devonian crustal-scale extensional movements. These L_2-lineations have orientations parallel to the F_3-folds, with plunge **around 10°** towards WNW (e.g. compare **Fig. 3.28** and **Fig. 3.31a**). This structural colinearity of L_2 and F_3 may be explained in two ways, either **(1)**: F_3-folding occurred during the D_2-shear movements, or **(2)**: the F_3-folding is related to a phase of deformation that entirely post-dates the D_2-shearing. As we shall see, these interpretations have fundamental consequences for the geological development of the area.

In alternative **(1)**, it is assumed that the folding of the mylonite fabric has occurred sometime *during* the extensional movements, possibly at the later stages of the shear movements. This development would be compatible to the "detachment" model, which assumes that transpressional forces, due to e.g. lateral ramps, could produce folding of the mylonites in the detachment zone (in addition to folding of the Upper Plate/Devonian basins), during the overall extensional movements (e.g. Hossack 1984; Norton 1987; Chauvet & Seranne 1989; Larsen 2002b). Various types of such syn-extensional folding is well documented in the literature, e.g. from the Basin and Range area of the USA (e.g. Yin 1991; Fletcher & Bartley 1994, Fletcher et al. 1995; Mancktelov & Pavlis 1994; Janecke et al. 1998) and from other places around the world (Schlische 1995).

In alternative **(2)**, the colinearity might have been a coincidental result of two indipendent stress systems, starting with a phase of crustal-scale E-W extension (D_2), and succeeded by a later phase of N-S shortening (D_3). This model is also compatible with the "detachment" model (see Sect. 2.8), if it is assumed that large-scale Devonian sinistral wrench tectonics (to the NW of western Norway) first controlled the E-W extension, but at a late stage led to N-S contraction (cf. Seranne et al. 1991). The N-S contraction as such, might also give the impression of complying with the "Solundian Orogeny" model (see Sect. 2.9), but other major aspects of the "Solundian Orogeny" model make this interpretation problematic. See discussions in Ch. 6.

Final remarks on the structural development of the Eikefjord Group

The age relationships between the L_2- and the F_3-structures are of great importance in the interpretation of the time and cause of the F_3-folding. The S_2-mylonites and the L_2-lineations in the Eikefjord Group rocks formed during extensional movements, and the fabric has been interpreted as part of the crustal-scale Nordfjord–Sogn Detachment Zone. Major extension on this zone led to the formation of the Devonian basins on the Upper Plate in the west. This means that F_3-folding of the S_2-mylonites have occurred at least *after the onset* of these extensional movements, possibly at the *late stage of*, or *after*, the shearing. The tectonic processes that produced the F_3-folds may alo have caused the folding of the Devonian massifs, and the folding may therefore have occurred during or after Middle Devonian times. These issues are further discussed in Ch. 6.

3.5 METAMORPHISM

The metamorphic data presented here are taken from some of the representative rock types described in Sect. 3.3. The analysis of the metamorphic development of the rocks is based on thin-section studies of mineral assemblages and textures; geobarometry/-thermometry or geochemical analyses have been beyond the scope of this thesis. Section 3.5 starts with a brief consideration of the possibilities for Precambrian metamorphic signatures (Sect. 3.5.1), but the main purpose of the present account is to unravel the metamorphic development during the extensive D_2-mylonitisation (Sect. 3.5.2). Metamorphic conditions during F_3-folding has been more difficult to establish due to lack of index minerals. The metamorphic development of the Eikefjord Group is interpreted as a continuos process, and therefore notations like M_1, M_2, etc. have not been used.

3.5.1 POSSIBLE PRECAMBRIAN METAMORPHISM

The meta-anorthosites and related rocks of the Eikefjord Group are generally considered to belong to an association of rocks which has been termed the "Anorthosite-Jotun-Kindred" (Goldschmidt 1916, Bryhni et al. 1981). These rocks are present in several areas in western South Norway, but the three main areas, apart from the Eikefjord–Gloppen area (cf. Sect. 2.3), are represented by the Dalsfjord Suite (Dalsfjord Nappe) (cf. Sect. 2.5), the Jotun Nappe Complex, and the Lindås Complex near Bergen. In these three areas, the rocks have, to a variable extent, preserved an inherited granulite facies mineral assemblage of Precambrian age as evidence of their oldest metamorphic history (e.g. Schärer 1980; Corfu & Andersen 2002; Austrheim & Griffin 1985; Lundmark et al. 2007). The later, Caledonian overprinting is generally confined to greenschist facies for the Dalsfjord and Jotun Nappes, and amphibolite facies for the Lindås Nappe, but relicts of the Precambrian high-grade assemblages have frequently survived this recrystallisation (Bryhni and Sturt 1985; Milnes & Koestler 1985; Austrheim & Griffin 1985).

In the investigated part of the Eikefjord Group, thin-section studies of the mylonites have not revealed unequivocal evidence of relict granulite facies mineral assemblages. However, from other parts of the Eikefjord Group, Bryhni et al. (1981) reported relict string perthites and ortho-pyroxenes that were interpreted as remnants of the granulite facies. In addition, Bryhni (1958) also reported lenses with undeformed anorthosite near Knapstad, **2 km** east of Eikefjord village, where the rock contained coarce-grained (diametres up to **several cm**) violet-coloured plagioclase of labradoritic composition (**An 50–60%**). This plagioclase composition was interpreted to show that the rock had preserved its Precambrian granulite facies mineralogy.

Feldspars with microscopic exolution lamellae ("perthites") are commonly present in the rocks studied in the present thesis, especially in the syenitic to monzonitic grey gneisses (**Fig. 3.10**), but it is uncertain whether these perthites are really granulite facies variants of the same type as those reported by Bryhni et al. (1981), or a lower grade variant which is fairly common in granitoid igneous rocks (pers. com. Kjell-Petter

Skjerlie). Ortho-pyroxene relicts have not been observed. The apparent lack of conclusive granulite facies relicts in the studied area is probably a consequence of the extensive reworking during the mylonitisation process, and does not exclude the possibility that such relicts might be present elsewhere in the Eikefjord Group.

3.5.2 POST-SCANDIAN DEVONIAN METAMORPHISM

Metamorphic minerals and textures related to the D_2-mylonitisation

The metamorphism in the rocks of the Eikefjord Group is closely related to the penetrative D_2-mylonitisation process. The S_2-mylonitic foliation and related stuctures were described in Sect. 3.4. The metamorphic state of the rocks during mylonitisation can be illustrated as follows:

In the ***meta-anorthosites***, the plagioclase minerals, which generally have been recrystallised during the mylonitisation, have an *albitic* composition. As mentioned above, Bryhni (1958) reported that prior to mylonitisation, the plagioclase had a labradoritic composition (**An 50–60%**). The plagioclase in the study area has been partly altered to sericite, and **up to 15 %** sericite may be present. Epidote and zoisite can constitute **up to 25 %** of the rock, indicating that the An-content of the original plagioclase has been strongly reduced. Retrogressive chlorite is part of the mylonite fabric, having grown partly from epidote and zoisite. Chlorite, epidote, zoisite and white mica can be evenly distributed among the plagioclase subgrains (**Fig. 3.5**). The chlorite content can be **up to 15–20 %**. The *grey gneiss* has a relatively high content of epidote and biotite (**Fig. 3.9**). Retrogressive chlorite is present in the mylonite fabric, and has been partly growing from epidote and biotite. The plagioclase subgrains have an albitic composition. The *amphibolite* contains chlorite (and actinolite) as retrogressive products of hornblende and biotite (**Fig. 3.19**). In the *micaschist*, chlorite is again part of the mylonite fabric. Garnet has been retrograded to chlorite and biotite (**Fig. 3.20**). The plagioclase is albitic.
The *meta-gabbros* have been reported by Bryhni (1958) to show extensive retrogression to an actinolite-chlorite-epidote rock.

On the microscopic scale (and also on outcrop-scale), bands of chlorite are anastomosing through the mylonitic plagioclase fabric in various rocks, and can be particularly well displayed in the meta-anorthosite (**Fig. 3.23**). The chlorite occurs in shear bands within the mylonite fabric and constitute an integral part of it. The plagioclase grains approach a semi-brittle behaviour, and the chlorite bands isolate lense-shaped plagioclase grains/aggregates. On outcrop-scale, porphyroclasts of feldspar may show cracking (**Fig. 3.16**), and cracks may also be present between "shear surfaces" within the rock itself (**Fig. 3.25**). Such brittle behaviour of feldspar suggests temperatures **below 500 °C** (Vidal et al. 1980; Olsen & Kohlstedt 1985; Tullis & Yund 1987; Pryer 1993). The chlorite in the shear bands is interpreted to have grown *syn-kinematically*, possibly during the last stages of top-to-the-west shearing. Post-kinematic *mimetic* growth of chlorite also appears to be present. This last process may indicate that the rock volume as a whole has attempted to adjust to shallow crustal metamorphic conditions, and the growth may have been facilitated by fluids migrating through the NSDZ.

Conclusion on the degree of metamorphism during D_2-mylonitisation

From the above description, it appears that the rocks show a retro-metamorphic, syntectonic recrystallisation going from hornblende, zoisite/epidote, garnet and Ca-plagioclase; to biotite, actinolite, albite, white mica and chlorite. The “highest” part of the mineral assemblage is equivalent to the ***amphibolite facies***. (Turner 1981; Yardley 1989). In the mylonitic anorthosites, the occational presence of a texture that is dominated by large ductily (plastically) extended feldspars indicates that this metamorphic level has been present. Also the extensive recrystallisation of the large feldspar grains into subgrains and tiny newgrains, required amphibolite facies, although of a lower grade within this metamorphic facies. The various amphibolite facies minerals, however, show extensive alteration to minerals in the “lowest” part of the paragenesis listed, implying retrogression down to at least medium or probably lower *greenshist facies*. This is evidenced, for example, by the strong chloritisation (Turner 1981; Yardley 1989), and brittle behaviour of feldspar.

It is significant to note that the plagioclase that has been recrystallised into mosaics of tiny new-grains, in the mylonitic anorthosite, has an *albitic* composition (Bryhni 1958). As indicated above, such extensive shear-related recrystallisation of plagioclase (into sub-grains and new-grains) is a result of crystal plastic deformation, which for this mineral indicates at least amphibolite facies conditions. However, the fact that the recrystallised plagioclase has an *albitic* composition suggests greenschist facies environments, and appears to indicate that the latest shear movements within the mylonitic plagioclase fabric occurred during greenschist facies conditions, whilst preserving or reproducing the strongly recrystallised mylonitic texture.

As noted above, the brittle behaviour of feldspar, which possibly occurred during the later stages of the shear movements, is compatible with greenschist facies P/T conditions. Although such brittle behaviour of feldspar could also in part be due to increased strain-rates, etc., diaphtoretic processes are considered to be the most likely explanation, since the feldspar-craking is accompanied by chloritic shear bands.

Growth of chlorite is, as indicated above, found to be both syn-tectonic, i.e. related to the top-to-the-west shear movements, and partly also post-tectonic, related to mimetic processes probably occurring just after the movements. For comparison it may be noted that the NSDZ-segment located just east of the Solund Devonian Massif (Hyllestad region) was reported to contain syn-tectonic growth of chlorite in top-to-the-west-related shear bands (Chauvet & Seranne 1989) (see **Fig. 2.3** for location). From the same area, mimetic growth of chlorite was reported by Chauvet et. al. (1992).

Wilks & Cuthbert (1994) investigated the metamorphic development of the Eikefjord–Gloppen area, and concluded that most of the shear movements occurred under amphibolite facies conditions. This was based on abundant presence of sheared plagioclase, showing recovery-accommodated crystal plastic deformation textures (e.g. subgrain rotation), features that, with referrence to to Tullis & Yund (1985, 1987), were taken to indicate deformation in the middle amphibolite facies or higher. Wilks & Cuthbert (1994) further reported that greenschist facies mineral assemblages were rarely encountered, but, when present, were limited to retrogression of amphibolite facies assemblages, and usually confined to extensional crenulation cleavages. Such cleavage surfaces were reported to be associated with **(i)** extreme grain size reduction of quartz and biotite, **(ii)** a change in biotite colour from khaki to pale green, **(iii)** growth of chlorite, and **(iv)** occationally adjacent garnet grains being partially replaced by chlorite. Also Johnston et al (2007b) suggested that most of the shear occurred under amphibolite facies conditions.

Johnston et al. (2007b) studied the metamorphic development of the Eikefjord rocks (and the Lykkjebø rocks), in the Eikefjord–Hyen area. Thermobarometric analyses indicated that the Eikefjord and Lykkjebø Groups both experienced prograde upper amphibolite facies peak metamorphism at **13–18 kbar, 537–618 °C**, **45–60 km** depth, interpreted to reflect conditions prior to the development of the NSDZ, followed by retrograde lower amphibolite to greenschist facies metamorphism at **8–12 kbar, 519– 641°C**, **30–40 km** depth, interpreted to be related to top-W shear movements on the NSDZ. Also These authors suggested that a major part of the movements occurred during amphibolite facies conditions.

However, the view of Wilks & Cuthbert (1994) and Johnston et a. (2007b) that greenschist facies assemblages are less prominent in these rocks, cannot be confirmed by the present work. On the contrary, and in accordance with Bryhni (1958) who made detailed petrographic studies of the area, the retrogression-effects and presence of greenschist facies assemblages in the rocks are extensive, occurring in localised shear bands, generally in the shear zone, and also throughout larger lenses/rock bodies that have escaped extensive shearing. Consequently, a large portion of the movements occurred during greenschist facies conditions. Due to rheological factors it cannot be expected that shear-related fabrics under greenschist facies conditions will develop as penetratively as in the amphibolite facies. The shear will tend to be more localized, and feldspar-rich rocks will be particularly unfavourable for penetrative shear.

The rocks appear to have reached the greenscfist facies by retrogressive processes. Although a *prograde* (instead of a retrograde) P/T development into the greenschist facies could theoretically produce *retrograde* mineral transformations and textures, this possibility is unlikely in the present case. Such a metamorphic development would imply that the rocks had experienced burial, and not exhumation. The range of retrograde mineral reactions described above, combined with a complete lack of signs of *prograde* mineral reactions, disfavours the possibility that the retrograde mineral reactions is a result of a *prograde* P/T development.

In summary, the rocks thus contain a *complete range* of metamorphic minerals from amphibolite facies assemblages to greenschist facies assemblages, and it is suggested that the minerals reflect a *continuous* retro-metamorphic development. Such a retrogressive development of the NSDZ has been reported from several other segments of the NSDZ, notably east of the Solund Devonian Massif (Tillung 1999; Hacker et al. 2003), and below the Kvamshesten massif (Swensson & Andersen 1991). This suggests that the minerals, during mylonitisation, continuously experienced more shallow P/T conditions, i.e. experienced unroofing. Despite the high shear strains that characterise a large shear system as the NSDZ, the minerals did not achieve recrystallisation to equilibrium conditions in the greenschist facies.

The present results on retrogressive metamorphic development confirm earlier investigations in the local NSDZ segments, notably in the Eikefjord Group, described by Bryhni (1958); and in analoguous mylonites along the Standalen Detachment to the south of the Håsteinen Devonian Massif, studied by Andersen & Jamtveit (1990).

Based on the present knowledge, therefore, it seems likely that the mineralogical changes are related to metamorphism during *one* continuous unroofing process, giving gradual retrogression from amphibolite facies to greenschist facies.

Interpretation of the S_2-related retrograde metamorphism

The rocks of the Eikefjord Group were in Sect. 3.4.2 interpreted as located within the Nordfjord–Sogn Detachment zone (NSDZ), and the S_2-mylonitic fabric was ascribed to the top-to-the-west shearing in this zone. The metamorphic retrogression of these rocks from amphibolite facies to greenschist facies is interpreted as related to the crustal-scale unroofing that occurred during the extensional movements on the NSDZ. The shear-related amphibolite facies assemblage represents the highest metamorphic grade documented in the study area, but the age and origin of the amphibolite facies assemblage itself has been more uncertain. The following discussion gives a review of published $^{40}Ar/^{39}Ar$ ages from the Nordfjord–Sogn area, and then Sm-Nd ages recently published by Johnston et al. (2007b) from areas adjacent to Eikefjord.

$^{40}Ar/^{39}Ar$ datings of hornblende

$^{40}Ar/^{39}Ar$ datings of hornblende that have been reported in the literature may provide constraints on the tectonometamorphic development during lower amphibolite facies, since the hornblende ages indicate the timing of exhumation of the rocks to a crustal level shallower than the Ar-retention temperature of **500–550 °C** (Fossen & Dunlap 1998b, referring to Harrison 1981; Baldwin et al. 1990). Although none $^{40}Ar/^{39}Ar$ dating of hornblende has been reported in the literature from the *Eikefjord/Lykkjebø rocks* between the Håsteinen and Hornelen Devonian Massifs, the method has been applied to NSDZ- and WGR-rocks that are located north of the Hornelen massif, as well as between the Kvamshesten and Solund massifs — as follows:

---Lux (1985) conducted a $^{40}Ar/^{39}Ar$ hornblende dating in the WGR outside the NSDZ, at the little fjord Nordpollen located **13 km** north of the Hornelen Devonian Massif and **10 km** ENE of Måløy in outer Nordfjord, and obtained a plateau age of **410 +/- 1 Ma**.

---Chauvet & Dallmeyer (1992) carried out several $^{40}Ar/^{39}Ar$ hornblende cooling-age determinations:

(i) At Stryn, east of the Hornelen massif, the dating gave a meaningless age of **533 +/- 3.3 Ma** (sample H4) and a calculated IC age (Isotopic Correlation diagram age) of **414.8 +/- 6.2 Ma**, but suspicion of excess Ar was reported. The age will be ignored in the following treatment (as it was in Fossen & Dunlap 1998b).

(ii) In the NSDZ near Rugsund (less than **1 km** north of the Hornelen massif, and **ca 5 km** ENE of the very “Hornelen” mountain peak), a $^{40}Ar/^{39}Ar$ hornblende plateau age of **421 +/- 1 Ma** (sample H5) was obtained, which was also suspected of excess Ar.

(iii) At Hyllestad, within the NSDZ **ca 7 km** east of the Solund Devonian Massif, hornblende dating gave the plateau age **410.6 +/- 1.5 Ma** (sample H3), again suspected of excess Ar.

(iv) In the easternmost part of the Hyllestad region, at a locality **~ 4.5 km** ENE of Lavik, near Alværa, (i.e. just east of the NSDZ), hornblende dating yielded the plateau age **394.8 +/- 3.5 Ma** (sample H2), not suspected of excess Ar.

As noted by Fossen & Dallmeyer (1998), the above age spectra of Chauvet & Dallmeyer (1992) was markedly disturbed, suggesting only partial resetting of older mineralogies during Caledonian metamorphism at these locations. The ages from the three above localities (ii), (iii) and (iv) will be used below. (These ages were also used in the review by Fossen & Dunlap 1998b).

--- Eide et al. (1999) conducted a $^{40}Ar/^{39}Ar$ hornblende dating in the WGR south of the Kvamshesten Devonian Massif, i.e. outside the NSDZ, at a locality south of Dalsfjorden, near Folkestad south of Fure, and obtained a weighted age of **412 +/- 2 Ma**, again reporting effects of excess Ar.

Although some of the above datings were obtained from areas just outside the NSDZ, the sample localities are situated fairly close to the NSDZ, suggesting that they had an unroofing history not very different from that of the NSDZ. Within or close to the NSDZ itself, the three ages given by the **410.6 +/- 1.5 Ma** (iii) age at Hyllestad **7 km** east of the Solund massif (Chauvet & Dallmeyer 1992); the **410 +/- 1 Ma** age at Nordfjord (Lux 1985); and the **412 +/- 2 Ma** age south of Dalsfjorden (Eide et al 1999), of which the Nordfjord age was not reported to be suspected of excess Ar, form a group of overlapping ages centered around **~ 412– 410 Ma**. Although the **Dalsfjord 412 Ma** and the Hyllestad **410 Ma** hornblende ages were obtained from localities just outside the NSDZ, the unroofing history of the localities was probably not very different from that in the NSDZ.

Judged *solely* from the above hornblende datings, it may be suggested that the hornblende cooled through its retention temperature **500–550 °C**, at **~ 412– 410 Ma**, implying that the rocks experienced lower amphibolite facies metamorphism at this time.

However, when discussing the hornblende age of these rocks, some other factors should also be considered:

The first factor regards the issue of excess argon: As mentioned, the Dalsfjord and Hyllestad hornblende ages are reported to be suspected of excess argon, suggesting that the ages may be somewhat old. The second factor conserns the hornblende (number iv)-age of **394.8 +/- 3.5 Ma**, that was obtained from the easternmost Hyllestad region east of Solund, at a locality near Alværa **ca 4.5 km** ENE of Lavik: This age is **~ 15 Ma** younger than the other hornblende ages (of **~ 410 Ma**) referred above. In addition to the Hyllestad hornblende ages, four $^{39}Ar/^{40}Ar$ ***muscovite*** ages have been obtained from the Hyllestad region, of which three ages span the interval **404–399 Ma** (plateau ages, Chauvet & Dallmeyer 1992, samples no M3, M4, M5), and the fourth age was **395 +/- 1 Ma** (plateau age, Chauvet & Dallmeyer 1992, sample no M2). As we see, the young hornblende age of **394.8 +/- 3.5 Ma** is in fact *younger* than three *muscovite* plateau ages spanning **404–399 Ma**, and does *overlap* with the muscovite age of **395 +/- 1 Ma**, of which the last muscovite sample was taken from the easternmost part of the Hyllestad region, at a locality **ca 2 km** NW of Lavik (i.e. from a locality positioned **ca 6 km** WSW of the hornblende locality). In the light of the large areal dimensions of the outcropping NSDZ in the Hyllestad region, both the hornblende and *muscovite* sample localities near Lavik may be considered as situated roughly at the same structural position, i.e. just below and east of the NSDZ. Muscovite has an argon retention temperature of **350– 400°C** (Fossen & Dunlap 1998b, referring to Dodson 1973; Wagner et al. 1977; Snee et al. 1988; Hames & Bowring 1994), which is **100–200 °C** lower than the hornblende retention temperature of **500–550 °C**. In essence, the comparison of the four Hyllestad muscovite ages on the one hand, with the single easternmost Hyllestad hornblende age on the other, suggest that:

- the muscovite localities passed the **350– 400 °C** interval at **404–395 Ma**, whereas
- the eastern hornblende locality passed the **500–550 °C** interval at **~ 394.8 +/- 3.5 Ma**,

i.e. indicating a hornblende age that is either overlapping with or younger than the muscovite age interval.

These two, relatively closely neighbouring localities of hornblende- and muscovite-dated rocks in the Lavik area cannot have experienced the two respective retention temperature levels simultaneously, and it is therefore possible that the hornblende age of **394.8 +/- 3.5 Ma** is too young. This is indicated by the fact that the hornblende age stems from a locality positioned furtherst to the *east* among the Hyllestad region-ages reported by Chauvet & Dallmeyer (1992), a locality which should rather have been uplifted across the hornblende retention temperature boundary of **500–550 °C** at an *early* point in time during the exhumation of the Lower Plate, and thus should have yielded *old* hornblende cooling ages. The possibility that the hornblende age of **394.8 +/- 3.5 Ma** is too young is also indicated by the fact that the four *muscovite* ages of the Hyllestad region is consentrated in the age range **404–395 Ma**.

The above argument, that the rocks cannot have experienced hornblende and muscovite retention temperatures simultaneously, is of cource based on the condition that the neighbouring rocks were located relatively *close* also during their exhumation. This condition could be questioned, as it may be argued that the continued extensional movements in a mega shear zone like the NSDZ, could in fact bring rocks — that were at e.g. the **~395 Ma** point in time positoned at different crustal levels — to juxtaposed positions on a later stage. In such a case, the **394.8 +/- 3.5 Ma** hornblende age would not necessarely be too young.

Nevertheless, when considering the above factors, the consentration of Hyllestad muscovite ages in the interval **404–395 Ma** appears desicive, thus implying that the **394.8 +/- 3.5 Ma** hornblende age from near Lavik at easternmost Hyllestad is most likely too young. This interprertation is furthermore supported by the fact that the muscovite ages in NSDZ segments further north are concurrent with those of the Hyllestad segment; for example in the NSDZ segment between Håsteinen and Hornelen, where the muscovite ages fall in the interval **404–393 Ma** (see below).

Since the retention temperature of *muscovite* in the $^{40}Ar/^{39}Ar$ method is lower than that of hornblende, the hornblende ages should be expected to be statistically slightly *older* than the muscovite age, at least when the two dating methods are applied to samples from nearby situated localities. The datings discussed above show that the hornblende ages are indeed older than the muscovite ages, albeit with exception of the **394.8 +/- 3.5 Ma** hornblende age.

As indicated above, several of the hornblende ages are reported to be suspected of excess argon. This implies that the hornblende ages are possibly a bit too old, suggesting that the ages should have been somewhat younger than the **~ 412– 410 Ma** indicated above. The fact that the *muscovite* ages in the three NSDZ segments of the Solund (Hyllestad); Kvamshesten; and Håsteinen–Hornelen regions fall in the interval **404–393 Ma**, indicates that the majority of *hornblende* ages should be slightly older than this, although some degree of overlap with the muscovite ages could be expected. It is possible that the retention temperature of **500–550 °C**, that defined the hornblende age of **~ 412– 410 Ma**, could have lasted until for example **~ 405 Ma**, i.e. approaching the muscovite age interval **404–393 Ma**.

From the above discussion of hornblende ages as well as the referred muscovite ages, it may be suggested that the hornblende of the NSDZ started to cool through its retention temperature of **500–550 °C** at **~ 412– 410 Ma,** and that parts of the Hyllestad rocks possibly continued to experience these temperatures until **~ 405 Ma**, implying that the rocks experienced lower amphibolite facies metamorphism at this time.

In conclusion, it is therefore considered most likely that the hornblende minerals experienced the temperature of **500–550 °C,** in the interval **~ 412– 405 Ma**.

^{40}Ar/^{39}Ar dating of muscovite

^{40}Ar/^{39}Ar dating of muscovite from areas adjacent to the study area, has been reported in the literature, notably from the Eikefjord- and Lykkjebø rocks between Håsteinen and Hornelen, as well as east and north of Hornelen, and south of Håsteinen. As mentioned above, the method gives the point in time when the rocks were exhumed to temperatures lower than the retention temperature of **350–400 °C**. This temperature would correspond to lower greenschist facies conditions, which has been suggested above to be the degree of metamorphism present during the last stages of the shear movements in the Eikefjord Group.

Chauvet & Dallmeyer (1992) obtained the following ^{40}Ar/^{39}Ar muscovite ages:

393.0 +/- 0.7 Ma, ca **23 km** east of Hornelen, at northern end of lake Breimsvatnet (sample M6),

402.9 +/ 1.1 Ma, between Håsteinen and Hornelen, at Lykkjebø **4 km** north of Storebru, along the road Storebru–Hyen (sample M7),

399.2 +/- 0.7 Ma, some **hundred metres** south of Hornelen, at a locality **4 km** east of Haukå, along the main road (sample M8).

Andersen (1998) obtained the following ages:

396 Ma, 2.7 km east of Hornelen, at the eastern shore of central part of Hyenfjorden,

415 Ma, 2.7 km SE of Hornelen, at the north end of Emhjellevatnet, along the road Storebru–Hyen,

416 Ma, between Håsteinen and Hornelen, **7 km** south of Hornelen, at the northern shore of Krokstadvatn, along the road Storebru–Hyen,

404 Ma, between Håsteinen and Hornelen, **5 km** north of eastern part of Håsteinen, at

Storbru, the road junction between the Storebru–Hyen road and route 5 Florø–Førde.

399 Ma, between Håsteinen and Hornelen, near Svarthumle along route 5 Florø–Førde, **ca 11 km** east of Eikefjord,

398 Ma, **7.5 km** WSW of Håsteinen, at Standal, where the Standal Fault reaches the fjord.

In the paper of Andersen (1998), the ages were presented only in the form of age numbers printed on a map-figure that shows the Nordfjord–Sogn region, with the numbers positioned near the respective sample sites. Apart from the ages themselves, no other details of the datings were given. The two ages of **415** and **416 Ma** must be considered as anomalously high compared to the others, and will be ignored here. The other ages of the Eikefjord–Gloppen area fall in the interval **404–393 Ma**, which corresponds to the age when the rocks passed the retention temperature of **350–400 °C** during exhumation.

North of Hornelen, in the Nordfjord district, the numerous $^{40}Ar/^{39}Ar$ muscovite datings in the NSDZ and WGR (Cuthbert 1991- K/Ar dating; Chauvet & Dallmeyer 1992; Andersen 1998) lie in the interval **390–385 Ma**, i.e. substantially younger than the Eikefjord–Lykkejbø rocks ages of **404–393 Ma**. As noted above, the ages from east of the Solund massif (Chauvet & Dallmeyer 1992), that fall in the interval **404–395 Ma**, overlap with the Eikefjord–Lykkjebø ages.

***Note of caution*:** Carswell et al. (2003a, 2003b) reported that the UHP metamorphism of Hareidlandet SW of Ålesund, WGR, occurred at **402 +/- 2 Ma** (U/Pb on zircons), indicating that this was the time when the rocks were subjected to *fluids* that triggered and facilitated the metamorphic reactions producing the HP and UHP eclogite mineralogies (Austrheim 1998; Carswell et al. 2001; Engvik et al. 2000; Wain et al. 2001). Although it is not possible to conclude with certainty that the age represent the timing of maximum P and T (i.e. maximum depth) for the subducted WGR rocks, Carswell et al. (2003a) nonetheless suggested that the maximum P and T would (quote) “conceivably [lie] within the precision of the determined U/Pb zircon age”. A new dating of UHP eclogites north of Hornindalsvatnet in Nordfjord gave the age **405 +/- 2 Ma** (Young et al. 2007), i.e. in the same range as the age of Carswell et al. (2003a, 2003b). Anyway, the age of the UHP metamorphism appears to contradict the $^{40}Ar/^{39}Ar$ hornblende dating of Lux (1985), which states that the rocks in the vicinity of the Stadtlandet area had been exhumed to lower amphibolite facies at **410 +/- 1 Ma**, i.e. **ca 8 million years** earlier than the UHP age of **402 +/- 2 Ma**. In their paper, Carswell et al. (2003b) solved this appearant contradiction by generally stating that the $^{40}Ar/^{39}Ar$ dating method on hornblende and micas is an unreliable method due to excess argon. When we consider the mineral *hornblende*, the $^{40}Ar/^{39}Ar$ datings of this mineral in the Sogn–Sunnfjord area and WGR of western Norway, do indeed show a certain spread, which may in fact be due to excess argon. These hornblende datings should therefore be treated with caution (this work). The *mica ages*, however, are very consistent within districts of the Sogn–Nordfjord region, suggesting that the ages are more reliable.

Carswell et al. (2003b) furthermore suggested that the eclogites were exhumed to a depth of **~ 35 km** (i.e. still high P/T, deep crustal level) by **395 Ma**, i.e. **ca 7 million years** later than the **402 +/- 2 Ma** eclogite age, implying an exhumation rate of **10 mm a^{-1}**. The **395 Ma** age is **5 million years** earlier than the $^{40}Ar/^{39}Ar$ muscovite ages of **390–385 Ma** north of Hornelen. The area investigated by Carswell et al. (2003b) is located in the part of the WGR that was subducted to the deepest level. It is therefore reasonable to assume that the NSDZ further south (e.g. in the Eikefjord–Gloppen area), as well as the adjacent/subjacent WGR, were exhumed to shallow crustal levels earlier than the UHP areas of Stadtlandet–Ålesund. The fact that the mica datings north of Hornelen (see above) fall in the range **390–385 Ma**, therefore appears to be fully compatible with the UHP dating and exhumation history suggested by Carswell et al. (2003b).

In summary, it appears that the Eikefjord–Lykkjebø rocks and their D_2-fabric were probably exhumed through the retention temperature of **500–550 °C** at ca **412–405 Ma**, representing lower amphibolite facies, and through **350–400 °C** at **404–393 Ma**, representing lower greenschist facies. The greenschist facies minerals were clearly formed from retrogression of the amphibolite facies minerals.

Sm-Nd dating of garnet

Johnson et al. (2007b) performed Sm-Nd garnet dating of the Lykkejbø Group rocks, and found garnet core ages of **425–422 Ma** and rim ages of **415–407 Ma**. These ages were interpreted ro reflect continuos upper amphibolite facies conditions (**45–60 Km**) that existed prior to the formation of the NSDZ and that ended by **415–407 Ma**.The development of the NSDZ at lower amphibolite–greenschist facies was believed to have started subsequently, i.e. from around **410 Ma** (the mean value of the **415–407 Ma** interval). Since the movements along the zone were assumed to have occurred mainly at amphibolite facies conditions, the $^{40}Ar/^{39}Ar$ muscovite ages, reflecting retention temperature of **350–400** °C (lower greenschist facies) were taken to mark the end of the movements along the zone. Consequently the authors suggested that the movements on the NSDZ occurred in the interval **410–400 Ma**.

Age relationships to Mode-I

The time relationship between the age intervals given above and the formartion of the NSDZ as a ***Mode-II*** zone is of interest, since the datings are from samples taken within or near the mega shear zone. According to Fossen & Dunlap (1998), the preceding ***Mode-I*** extension that was defined by westward movement of the entire orogenic wedge along the subjacent décollement zone, occurred in the time interval **402–394 Ma**. During this time interval, the eclogitic Lower Plate was constantly exhumed, and the resulting uplift also of the **Mode-I** décollement zone finally rotated the zone to a subhorizontal orientation. This orientation eventually caused the zone to become mechanically locked due to increased friction, a situation that probably occurred around **394 Ma**, i.e. at the end of the "age interval" of the **Mode-I** movements. The locking of the **Mode-I** zone apparently led to formation of the nappe-transecting **Mode-II** NSDZ, and the formation of this zone presumably took place at the later stages of the **Mode-I** movements, i.e. some time before **394 Ma**. If this age for the formation of the NSDZ is correct, it implies that the hornblende ages of **~ 412–405 Ma** is clearly older than the formation of the NSDZ, and also implies that the muscovite ages of **404–393 Ma** are partly older and partly contemporaneous with the formation of the NSDZ. Movements at **410–400 Ma,** as suggested by the Sm-Nd ages of Johnston et al. (2007b) would also happen prior to the formation of Mode-I.

The amphibolite facies metamorphism, which apparently was related to Devonian metamorphic retrogression associated with top-to-the-west extension and unroofing, could itself, however, represent merely a stage in a continuous unroofing/retrogression process. Alternatively, the amphibolite facies metamorphism could have been related to the Scandian top-to-the-*east* contraction. These topics are further discussed in Sect. 6.4.4.

Metamorphic conditions during the F_3-folding

Metamorphic conditions during the F_3-folding is more uncertain. A D_3-related syntectonically recrystallized S_3-axial plane cleavage has not been observed, and metamorphic index minerals for the F_3-folds are thus lacking. The folding is, however, clearly ductile (crystal plastic), indicating that *minimum* P/T conditions might have corresponded to at least the lower greenschist facies. The lack of renewed mineral growth related to the dynamics of the F_3-folding could suggest that also a *maximum* P/T estimat could correspond to the greenschist facies, since higher P/T conditions (amphibolite facies) would more easily facilitate a D_3-related recrystallisation, particularly in the tightest F_3-folds. It can thus be assumed that the F_3-folding occurred under approximately the same P/T conditions as the late stages of the D_2-mylonitisation.

Comparison with metamorphism in the adjacent mylonites of the Høydalsfjorden Complex

The Lower Palaeozoic rocks of the Høydalsfjorden Complex, which will be discussed in more detail in the next main chapter (Ch. 4), have experienced a twofold tectonometamorphic development; (i) first the rocks were obducted and thrusted towards the east during the contractional Scandian phase, producing S_1-foliation and F_2-folds with a related M_1- and M_2-metamorphism (Sect. 4.4), and (ii) subsequently the rocks were effected by Devonian westward extension, producing S_3-mylonites, and possibly F_4-folds, with a related M_3-metamorphism (Sect. 4.4). The metamorphic conditions during the (i) contractional top-to-the-east event was different from the (ii) extensional top-to-the-west event, as briefly outlined in the following:

(i): As will be described in Sect. 4.4, the M_1- and M_2-metamorphism of the Høydalsfjorden rocks were *prograde* into the upper part of the *lower* greenshist facies. The mylonites of the Eikefjord Group, however, are *retrograde* well into the greenschist facies. Despite the different "directions" of approach towards the greenschist facies, it is worth noting that the metamorphism of the rocks in both groups approach the same P/T conditions. However, since the M_1- and M_2-metamorphism of the Høydalsfjorden rocks are related to the contractional top-to-the-east emplacement of the rocks during the Scandian continent-continent collision (Sect. 4.4), this metamorphism of the Høydalsfjorden rocks cannot be related to the top-to-the-west metamorphism of the Eikefjord Group.

(ii): The Høydalsfjorden rocks situated immediately south of the Eikefjord Group, i.e. south of the Sunnar fault, contain S_3-mylonites that were formed during top-to-the-west shear (Sect. 4.4.4). This shear direction appears to correspond to the similar direction in the Eikefjord Group. The Høydalsfjorden S_3-

mylonites define zones present in the mountainside rising from Sunnarviken and up to the rocks of the Håsteinen Devonian Massif. The S_3-mylonites formed at temperatures of ca **300 °C** (Sect. 4.4.4), reflecting metamorphic conditions just *below* the greenschist facies. The metamorphism during extensional movements in the Eikefjord Group corresponds to the middle or lower greenschist facies. Hence, the metamorphic conditions in the Høydalsfjorden mylonites were of a lower grade than those of the Eikefjord Group. The Eikefjord S_2-mylonites thus reflect a deeper crustal level than the Høydalsfjorden S_3-mylonites, a feature which is consistent with the higher tectonostratigraphic position of the Høydalsfjorden Complex. As outlined in Sect. 4.3.7.2, the Høydalsfjorden S_3-mylonites may have formed simultaneously with the *early* stages of the Eikefjord Group S_2-mylonites. Later, however, whilst the Høydalsfjorden S_3-mylonites were cut by the sub-Devonian unconformity, the Eikefjord S_2-mylonites had a continued, long-lasting story of movements during the Middle Devonian, that controlled the deposition of the Devonian basins.

Comparison with metamorphism in the Håsteinen Devonian Massif

As will be shown i Ch. 5, sandstones in the western part of the Håsteinen massif has been folded into roughly E-W trending folds that exhibit an axial planar cleavage. The P/T conditions during formation of this cleavage corresponded to the anchizone to lowermost greenschist facies. P/T conditions favourable for cleavage formation in the Håsteinen basin probably existed for a long period during the basin history, but the formation of the cleavage occurred during, and as a result of the N-S contraction of the basin. It is possible that this contraction was genetically connected to the N-S shortening that formed the F_3-folds in the Eikefjord Group. This will be further discussed in Ch. 6.

Time relationships between Devonian brittle faults at Øygarden (near Bergen), and movements on the NSDZ

In the Nordfjord–Bergen region, the ***early-stage mylonites*** (amphibolite facies) of the NSDZ rocks presumably formed prior to development of brittle extensional faults in the above-lying Upper Plate, faults that may have played a role in the formation of Devonian basins. These relationships seem to imply that the early-stage Eikefjord mylonites were formed prior to ORS deposition. A possible relationship between amphibolite facies mylonites in the NSDZ, and brittle faulting in the Upper Plate, may be visualised by a consideration of corresponding units south of the study area: south of the Kvamshesten Devonian Massif, as well as east of the Solund Devonian Massif, amphibole from amphibolite facies rocks in both the NSDZ and nearby situated rocks yielded an $^{40}Ar/^{39}Ar$ age of **~ 412–405 Ma** (see Sect. 3.5.2), indicating that the rocks passed the retention temperature of **500–550 °C** (Fossen & Dunlap 1998, referring to Harrison 1981; Baldwin et al 1990), i.e. lower amphibolite facies, at this time. Still further south, in the Øygarden Complex, which lies west of the Bergen Arc Shear Zone (Wennberg et al. 1998; see also Fossen & Hurich 2005), brittle faults contain mineralisations that gave a U/Pb sphene mean age of **396 Ma** (single ages spanning **399–393 Ma**), interpreted to date the fault formation (Larsen et al. 2003). Faults of similar age may be present in the Upper Plate between Bergen and Nordfjord, where such faults possibly influenced/controlled deposition of Devonian sediments. This may suggest a time gap of **~16–9 m.y.** between the **412–405 Ma** lower amphibolite facies of the NSDZ rocks and the possibly basin-controlling brittle faulting (of **396 Ma**) in the Øygarden Complex. A time gap would also be present between the NSDZ movement of **410–400 Ma,** suggested by Johnston et al. (2007b), and the brittle fault of **396 Ma**.

The *later-stage mylonites* of the NSDZ (lower/middle greenschist-facies) might have been contemporaneous with the development of faults that controlled formation of Devonian basins on the Upper Plate. This age relationship may be illustrated from the NSDZ-mylonites in the Håsteinen–Hornelen region. Here, greenschist-facies mylonites yielded $^{40}Ar/^{39}Ar$ muscovite ages spanning **404–398 Ma** (Andersen 1998; Chauvet & Dallmeyer 1992), indicating that the rocks passed the retention temperature of **350–400 °C** (Fossen & Dunlap 1998, referring to Wagner et al. 1977; Snee et al. 1988; Hames & Bowring 1994), corresponding to lower greenscist facies, at this time. The mean age of **~ 396 Ma** of the brittle faulting west of Bergen (spanning **399–393 Ma**, Larsen et al. 2003) overlaps with the above-mentioned lower greenschist facies age (of **404–398 Ma**) for the EG/NSDZ mylonites. The F_3-folds of the Eikefjord Group/NSDZ are folding the top-to-the-west Mode-II mylonite fabric of the rocks (**Table 4.1**).

Comment:

Mismatch between radiometric ages and the interpretation of the NSDZ as a Mode-II zone

$^{40}Ar/^{39}Ar$ amphibole ages (dating lower amphibolite facies in the mylonites)

As mentioned above, the dating method $^{40}Ar/^{39}Ar$ on amphibole, has a retention temperature of **500–550 °C** (Fossen & Dunlap 1998, referrring to Harrison 1981; Baldwin et al 1990), corresponding to lower amphibolite facies metamorphism. The **412–405 Ma** age ($^{40}Ar/^{39}Ar$ on amphibole, Chauvet & Dallmeyer 1992; Eide et al. 1999) of the top-to-the-*west* lower amphibolite facies mylonites of the Mode-II/NSDZ is time equivalent to the last part of the contractional Scandian phase of top-to-the-*east* movement of the orogenic wedge, which is documented (from the décollement zone below the Jotun Nappe) to have occurred in the time interval **415–408 Ma** (Fossen & Dunlap 1998). Nevertheless, it is, from field evidence, in the literature held that the westward Mode-I movements of the orogenic wedge (on the décollement zone) occurred first, whereas the Mode-II NSDZ formed, and moved, subsequently (Fossen 1992; 2000; Fossen & Dunlap 1998; Milnes et al 1997), albeit, with the Mode-II NSDZ presumably starting at the late stages of the preceeding Mode-I extension. Consequently, since the Mode-I extension has been dated to **402–394 Ma** (Fossen & Dunlap 1998), it is likely that the Mode-II NSDZ formed sometimes around **394 Ma**. However, if the NSDZ formed at this time, a mismatch would seem to arise between the **412–405 Ma** age of the lower amphibolite facies NSDZ mylonites (indicating top-to-the-west movements already at this time) and the status of the NSDZ as a Mode-II sone, the latter implying that the zone first formed around **394 Ma**, i.e. **11 Ma** later than **405 Ma**.

$^{40}Ar/^{39}Ar$ muscovite ages (dating lower greenschist facies in the mylonites)

The NSDZ mylonites in the Eikefjord–Hyen area yielded $^{40}Ar/^{39}Ar$ *muscovite* ages of **404–398 Ma** (Andersen 1998; Chauvet & Dallmeyer 1992), indicating that the rocks passed the retention temperature of

350–400 °C (Fossen & Dunlap 1998, referring to Wagner et al. 1977; Snee et al. 1988; Hames & Bowring 1994), corresponding to lower greenscist facies, at this time. Furthermore, it may be noted that this lower greenschist facies age interval of **404–398 Ma** for the EG ***Mode-II*** mylonites of the Håsteinen–Hornelen region overlaps with the ***Mode-I*** age of **402–394 Ma** that was obtained by Fossen & Dunlap (1998) from dated mica from samples subjected to the top-to-the-west ***Mode-I*** movements on the decollement zone below the Caledonian Jotun and Bergsdalen nappes. This would appear to indicate that the Eikefjord Group rocks had been exhumed to the **350–400 °C** temperature level, already during the Mode-I movements of the Caledonian nappes (**—** although one should keep in mind that the rocks below the Jotun and Bergsdalen nappes were never subjected to deep burrial/high-grade metamorphism of the type present in the lower part of the NSDZ, where the NSDZ are facing the eclogite-bearing Lower Plate).

Movements on the NSDZ as inferred from Sm-Nd garnet ages

Johnston et al. (2007b) suggested that the movements on the NSDZ occurred in the time interval **410–400 Ma**, i.e. subsequent to their peak upper amphibolite facies metamorphism ending at **~410 Ma**, and prior to the ^{40}Ar/^{39} Ar ages of **~400 Ma**. Also movements in the **410–400 Ma** interval would be early compared to the ***Mode-I*** age of **402–394 Ma** that was obtained by Fossen & Dunlap (1998).

Since the top-to-the-west mylonites of the **Mode-II**/NSDZ experienced lower *greenschist* facies conditions already at the time when the ***Mode-I*** *movements* occurred along the decollement zone, the preceeding extensive **Mode-II**/NSDZ *amphibolite* facies mylonites could probably not have formed in the time interval of the décollement Mode-I movements. It appears that this amphibolite facies portion of the shear history of the Mode-II/NSDZ may have taken place earlier, i.e. *before* the rocks of the Mode-II zone were exhumed to the lower greenschist facies crustal level, i.e. before the time interval **404–398 Ma**. If we choose to interpret the radiometric datings as recording westward *movements* in the NSDZ, the extensive history of amphibolite facies shear movements of the Mode-II/NSDZ would seem to have occurred *prior* to the Mode-I movements. In other words, the data could be taken to suggest that the order of the events were **not** Mode-I first and Mode-II next, but rather Mode-II amphibolite facies shear first, and then a coeval Mode-I and Mode-II greenschist facies shear, next. However, such a change in the order of the Mode-I and Mode-II events cannot be supported: the order of appearance of the two modes are independently determined from field relations (e.g. Fossen 1992; 2000; Milnes et al. 1997) which implies that the Mode-I did undoubtedly occur before Mode-II. Hence, there appears to be a mismatch between the ^{40}Ar/^{39}Ar datings of the NSDZ on the one hand, and the interpretation of the NSDZ as a Mode-II zone on the other. This appears to suggest that the ^{40}Ar/^{39}Ar ages cannot be used uncritically to explain the geological history of the area.

The causes of the mismatch are not obvious, but might theoretically be related to the dating methods themselves (wrong ages) or to wrong interpretation of the ages. Wilks and Cuthbert (1994) have interpreted the detachment mylonites between the Hornelen and Håsteinen Devonian Massifs to be of Mode-

I/décollement zone type, and not of Mode-II-type. This solution could explain the overlapping $^{40}Ar/^{39}Ar$ - *muscovite* ages of the NSDZ and the décollement zone below the Jotun Nappe, but the older lower amphibolite facies mylonite fabric of the NSDZ, dated by $^{40}Ar/^{39}Ar$ on amphibole to **~ 412–405 Ma**, would still appear to represent a mismatch.

A comprehensive treatment of the causes of the mismatch problem will be beyond the limits of the present work, but it is of interest to be aware of the problem. The author intends to address the subject in future works.

As mentioned above, field work has established that the movements along the Mode-I/décollement zone did undoubtedly occur prior to the formation of — and the movements along — the Mode-II/NSDZ (Fossen 1992; 2000; Milnes et al 1997). In the present work, the geological story will therefore comply with this relative age.

3.6 SUMMARY

In the study area, the Eikefjord Group consists of mylonitic meta-anorthosites, grey gneisses, meta-gabbros and amphibolites. Micaschists, which in fact belong to the Lykkjebø Group, has for simplicity been included in the Eikefjord Group. The rocks are part of the so-called "Anorthosite-Jotun Kindred" (Goldschmidt 1916; Bryhni et al. 1981), and are correlated with similar rocks in the Dalsfjord Nappe (Brekke & Solberg 1987; Corfu & Andersen 2002); ans in the Jotun Nappe Complex (Bryhni & Sturt 1985; Milnes & Koestler 1985; Lundmark et al. 2007), and in the Lindås Nappe (Austrheim & Griffin 1985).

The rocks are characterized by a strong mylonitic SL-fabric, which is related to a D_2-phase. The earlier structures have mostly been obliterated by this strong mylonitisation process, but remnants of such structures that are present outside the study area, are collectively assigned to D_1. The mylonitic S_2-foliation generally dips both to the south and north, although with a dominant dip towards south, and the L_2-stretching/mineral lineations have plunges gently towards W–WNW. Shear sense indicators in the mylonites show that the D_2-movement was top-to-the-west.

The rocks experienced metamorphic retrogression from amphibolite facies towards lower greenschist facies during the D_2-mylonitisation. This has been interpreted to show that the rocks continuously experienced shallower crustal levels, i.e. tectonic exhumation.

The S_2-mylonite fabric is folded by F_3-folds. Fold axes are oriented with an ESE–WNW trend and gentle westward plunge, and are parallel to the L_2-lineations. The folding is interpreted to have occurred *during* the shear movements, possibly at the later stages of the shear.

The S_2-mylonites in this area are interpreted to be part of the Nordfjord–Sogn Detachment zone, which in other parts of western Norway has been described by a number of authors (e.g. Michelsen 1986a; Norton 1987; Seranne & Seguret 1987; Chauvet & Seranne 1989; Andersen & Jamtveit 1990; Dewey et al. 1993; Wilks & Cuthbert 1994, who named it the “pricipal shear zone”; Krabbendam & Dewey 1998; Johnston et al. 2007b). The Eikefjord–Lykkjebø rocks were particularly studied by (i) Wilks & Cuthbert (1994) in the area between Eikefjord and Gloppen; (ii) by Andersen & Jamtveit (1990) from the NSDZ segment just to the south of the Håsteinen Devonian Massif; (iii) by Krabbendam & Dewey (1998) from the NSDZ-segement WSW of Håsteinen; and by (iv) Johnston et al. 2007b at Standal and between Eikefjord and Hyen.

The idea that the Eikefjord- and Lykkjebø Group rocks constitute a “Middle Plate”, first proposed by Andersen & Jamtveit (1990) and later applied by other authors, cannot be supported. The whole Eikefjord–Gloppen area is a penetratively mylonitised top-to-the-west detachment zone, as also thoroughly substansiated by Wilks & Cuthbert (1994) and further confirmed by Johnston et al (2007b). This shows that the area is part of the NSDZ. The metamorphic retrogression is interpreted to reflect the unroofing of subducted continental crust (Lower Plate) which occurred during the extension. The F_3-folding of the S_2-mylonite fabric may be interpreted to be a result of the same deformation process that deformed the Håsteinen Devonian Massif into a broad syncline. This is discussed in Ch. 6.

3.7 THE SUNNAR FAULT

The name "Sunnar Fault" was introduced by Bryhni et al. (1981, their Fig. 10 at p 24) to replace the name "Sunnar Line" (Bryhni 1958). The name “Sunnar Fault” has been changed to “Sunnarvik Fault” on the 1:50 000 maps of Bryhni & Lutro 2000a, 2000b, a name also adopted by Johnston et a. 2007b. However, the original name “Sunnar Fault” has been maintained in the present thesis. The fault runs southeastwards from Sunnarvika and meets the more prominent Standal Fault just to the east of the Håsteinen Devonian Massif (**Fig. 2.6**). A **2.6 km** long segment of the fault runs through the study area of the present thesis, between Sunnarviken and Vasset (**Plate 1**). The location of the structure can be observed on the main map (**Plate 1**).

Johnston et al. (2007b) assignet the fault to their group of “low-angle ductile.brittle detachments”, a group which was also stated to include the Hornelen detachment (below the Hornelen Devonian deposits) and the Standal detachment as a detachment. In the present thesis, the fault is rather viwed as a late, steeply dipping brittle structure.

The fault defines the contact between the Eikefjord Group to the north and the Høydalsfjorden Complex to the south. The fault contact itself is never completely exposed in the study area. To the west of a point located just south of the little lake Holatjønna (main map, **Plate 1**) the fault is mainly developed as a minor depression in the topography. To the east of this point, the Lower Palaeozoic rocks on the southern side of the fault commonly form an escarpment up to **4 m** high, offering rock exposures. On the north side of the fault, in this latter area, the Eikefjord Group rocks, being meta-anorthosites, are rarely exposed near the fault zone, and never at the fault itself. Elsewhere along the fault, the exposed rocks on each side of the contact are, at their closest, present only **a few metres** apart. On both sides of the contact, the rocks are generally mylonitic, and the fault truncates the mylonites.

The amount and direction of displacement are somewhat uncertain. Nevertheless, a relative down-movement of the Lower Palaeozoic rocks on the southern side is assumed, since these southern rocks are situated tectonostratigraphically on top of the Eikefjord Group, although being now absent on the northern side of the fault. For the same reason, it is likely that the amount of vertical separation is considerable, although this is uncertain. It is not known whether transcurrent movements have occurred.

Near the Vasset farm, a rock with a partly “cataclastic” appearance, is present for **several metres** along the escarpment on the southern side of the fault. The rock occurs in the steep cliff wall, but also on the flat *top* of the cliff, along its edge. Despite a low degree of exposure, the thickness of the rock may be estimated to at least **1 m**, possibly **2 m** or more. The colour of the rock varies from whitish on weathered surface to pale yellowish green on freshly broken surface. On an excavated surface, it was seen that the rock displays thin banding that was possibly formed by tectonic shear. On the same surface, outside the banded area, the rock also exhibited small, very angular rock fragments of mm- to cm-scale, that lay in a fine-grained structureless matrix consisting of a rock-type quite similar to the fragments themselves, giving the rock a “cataclastic” appearance. From its colour and “flint-hard” nature, the rock may be interpreted as an hydrothermal quartz-epidote “vein” that “intruded” along the Sunnar Fault. The fragments that create the cataclastic appearance may have formed as a result of either force-full “intrusion(s)”, or movements along the Sunnar Fault, followed by new pulses of vein fluids. Such hydrothermal quartz-epidote veins have been observed also elsewhere in the study area (e.g. at the island of Parisholmen in the westernmost part of the fjord Osstrupen, where a **10–15 cm** thick “vein” cuts conglomerate of the Håsteinen Devonian Massif, see Sect. 5.2.3). Quartz-epidote veins have also been described from the other Devonian massifs in western Norway (e.g. Noelle & Larsen 2000; Svensen et al. 2001; Larsen 2002b).

The “cataclastic” rock near Vasset has not been observed elsewhere along the Sunnar Fault, although this may be due to limited exposures. If it shold be correct, however, that the rock is present only at this particular point along the fault, and that the thickness of the rock is up to **several metres**, it might be speculated that the rock “body” represents a major feeder channel for hydrothermal fluids in the area. Alternatively, the considerable width of the rock at this point could have been achieved by repeated faulting combined with repeated fluid emplacement, thus adding to the thickness of the zone. This alternative would imply that the rock shold be present along large parts of the fault. The limited exposures, however, prevented a further clarification of these questions.

As mentioned in Sect. 2.3, Braathen et al. (2004) abandoned the name “Sunnar Fault” for this fault, and in stead called it the “Eikefjord Fault”. This is most unfortunate, as the name “Eikefjord Fault” has traditionally been used to denote the prominent E-W trending fault that defines the *northern* margin of the Eikefjord Group in the Eikefjord area (e.g. Bryhni et al. 1981; Bryhni & Lutro 1991a, 1991b; Berry et al. 1995: Krabbendam & Dewey 1998; Johnston et al. 2007b). As described in Section 3.2 above, the Eikefjord Fault *proper* passes E-wards from the public school building in the northern part of the village of Eikefjord, with a prominent cataclasite present in the up to **10 m** high cliff along the fault. The name “Sunnar Fault” (Bryhni 1981; Bryhni & Lutro 1991a, 1991b) (or the suggested new name “Sunnarvik Fault”, Bryhni & Lutro 2000a, 2000b; Johnston et al. 2007b) has traditonally been used to denote the fault at the *southern* margin of the Eikefjord Group, marking the contact between this group and the Høydalsfjorden Complex. To avoid confusion, the name “Eikefjord Fault” should not be transferred from the northern to the southern boundary of the Eikefjord Group. In the present thesis, the traditional use of the fault names has been maintained.

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Chapter 4

CHAPTER 4 HØYDALSFJORDEN COMPLEX (HC)

4.1 INTRODUCTION

Chapter 4 deals with the rocks of assumed Lower Palaeozoic age situated immediately around and beneath the Håsteinen Devonian Massif (**Fig. 4.1**, **Plate 1**). It is proposed in the present thesis that these rocks be assigned to a new complex termed the Høydalsfjorden Complex (HC).

Borders

To the north, the Høydalsfjorden complex are bordered by the brittle Sunnar Fault, which separated the rocks from the Eikefjord Group to the north. Note that the name "Sunnar Fault" has been changed to "Sunnarvik Fault" on the 1:50 000 maps of Bryhni & Lutro 2000a, 2000b, a name also adopted by Johnston et a. 2007b. However, the original name "Sunnar Fault" will be maintained throughout the present thesis. The southern border is defined by the brittle Standal Fault, to the south of which we find gneisses affected by the Nordfjord–Sogn Detachment Zone, and further south, the Westrrn Gneiss Region. To the east, the border is defined by the Sunnar and Standal Faults merging, causing the Høydalsfjord Complex to wedge out.

Naming

The name of the complex is taken from the fjord Høydalsfjorden located to the west of the study area. The rocks of the area have previously been named "*Heggøy Formation*" by Furnes et al. (1990), and "*Sunnarvik Group*" by Bryhni & Lutro (1991a, 1991b, 2000a, 2000b) and Johnston et al. (2007b), but for reasons explained below, it is here suggested that these names be abandoned.

The Heggøy formation of Furnes et al. (1990) was originally established on the island of Heggøy located west of the Kvamshesten Devonian Massif, where the Heggøy Formation was seen as part of the Stavenes Group that consituted the primary sedimentary cover sequence to the Solund–Stavfjorden Ophiolite Complex, S–SOC (see below). Based on reconnaissance work on road- and coastal sections along both sides of Høydalsfjorden, which revealed that the rocks consisted of mainly green- and grey-coloured greywackes — albeit intruded by numerous small gabbro bodies in a limited area west and northwest of the Håsteinen Devonian Massif, and possible felsic dykes locally around Høydalsfjorden — Furnes et al. (1990) assigned these rocks to their Heggøy Formation. However, although the greywackes clearly dominate on both sides of Høydals-fjorden, the work of the present author shows that the rocks continue eastwards into the study area where other rock types are present, for example meta-pelites, carbonates, organic-rich shales, large volumes of meta-psammites, etc. Also, outside the study area, particularly large meta-psammite bodies are present within the Høydalsfjorden Complex at Sandvika just northwest of the study area and on the islands of Ålvora and Stavøya in Høydalsfjorden. These features could suggest that the rocks are not merely a cover sequence to an ophiolite, but possibly a mix of continent-near deposits and “melange”-resembling elements (see below), hence justifying a separate name for the rocks — the Høydalsfjorden Complex.

The name “Sunnarvik Group” of Bryhni & Lutro (1991a, 1991b, 2000a, 2000b) and Johnston et al. (2007b), and also the name “Heggøy Formation”, are problematic due to violation of formal naming rules. The lithostratigraphic terms “group” and “formation” are not appropriate for the Høydalsfjorden Complex, since the unit contains a mixture of meta-sedimentary and meta-igneous rocks, and since there is no general control on the mutual stratigraphic positions of the meta-sedimentary rocks, i.e. no “way-up” control. Hence, the unit does not meet the requirements for lithostratigraphic naming established by the Norwegian Committee on Stratigraphy (Nystuen 1989), and therefore the litodemic term “complex” should be used instead.

When referring to the Devonian meta-sedimentary rocks that rest unconformably on top of the Høydalsfjorden rocks, the term “Håsteinen Devonian *Basin*” normally means the deposits as they were in the Devonian times or soon after, whereas the term “Håsteinen Devonian *Massif*” normally means the rocks as they appear today.

Regional correlation

Lower Palaeozoic meta-sedimentary and meta-igneous rocks that are similar to those in the study area, are present in the same tectonostratigraphic position in the area between the fjord Førdefjorden/Stavfjorden in the north and the fjord Dalsfjorden/Vilnesfjorden in the south, i.e. in the area to the west of the Kvamshesten Devonian Massif. These Lower Palaeozoic rocks are interpreted as being (i) part of either the Heggøy Formation, which is, as mentioned above, the primary sedimentary cover sequence to the Solund–Stavfjorden Ophiolite complex, S–SOC (Furnes et al. 1990), or (ii) part of the Sunnfjord Melange, which is an obduction melange present tectonostratigraphically just underneath the S–SOC (Andersen et al. 1990). Moreover, recent investigations of the Kalvåg Melange, which is an olistostromal (i.e. sedimentary) melange situated on the islands of Bremangerlandet and Frøya on the southern side of outer Nordfjord (Ravnås & Furnes 1995; Steen & Andresen 1997), have revealed features that may resemble features in the Høydalsfjorden Complex, hence opening for the possibility that parts of the Høydalsfjorden Complex rocks may be analogous to such an (iii) olistostromal (i.e. sedimentary) melange as well. It is therefore still an open question whether the rocks of the Høydalsfjorden Complex are in fact (i) part of a Heggøy Formation type sedimentary cover sequence (as tentatively suggested by Furnes et al.1990), or (ii) part of a Sunnfjord Melange type obduction melange, or (iii) part of a Kalvåg Melange type olistostromal melange (Harald Furnes pers. com.).

Age

The formation of the Solund–Stavfjorden Ophiolite Complex has been radiometrically dated by the U/Pb method at **443 +/- 3 Ma** (Dunning & Pedersen 1988). The age of the subjacent Sunnfjord Melange may be constrained as follows: The melange rests with a primary angular unconformity on the Herland Group, which represents Baltoscandian continental margin deposits. The Herland Group contains fossils of **Wenlock age (428–423 Ma**, time-scale of Gradstein & Ogg 1996; and Gradstein et al. 2004) and predates the obduction, suggesting that the obduction of the S-SOC started no earlier than **~425 Ma** (Andersen et al. 1990, Osmundsen & Andersen 1994; Milnes et al. 1997). The formation of the Kalvåg Melange is poorly dated by fossils to **Ordovician–Silurian** (Reusch 1903; Kolderup 1928), but Ravnås & Furnes (1995) interpreted the melange to have formed on a submarin volcanic arc-slope that bordered the marginal/back arc basin where the formation of the Solund–Stavfjorden Ophiolite Complex took place. All three units are thus believed to have been obducted/emplaced onto the Baltoscandian margin during the Scandian phase of the Caledonian Orogeny.

Purpose and structure of the chapter

The purpose of the present chapter is: (1) to give a short field description of the rocks, and (2) to present the main features of the tectono-metamorphic history of the rocks. A major objective has been to identify the tectono-metamorphic history that took place prior to the deposition of the Devonian rocks, i.e. to detect structures that are cut by the sub-Devonian unconformity. Due to limited exposures and lack of diagnostic features, the area investigated in the present work has not allowed firm conclusions to be drawn as to whether the rocks are part of a "sedimentary cover" to an ophiolite, an "olistostromal melange" or an "obduction melange", and the subject is therefore only briefly discussed.

Chapter 4 starts with a field description of both the meta-sedimentary and the meta-igneous rocks, including the intra-Devonian inliers, and briefly discusses possible environments for the formation of the rocks (Sect. 4.2). This is followed by an outline of the structural development, including a comparison with the structures in the Eikefjord Group (Sect. 4.3). Subsequently, the metamorphic development is briefly discussed (Sect. 4.4), followed by a summary and conclusions (Sect. 4.5).

4.2 ROCK DESCRIPTIONS

4.2.1 META-SEDIMENTARY ROCKS

4.2.1.1 GENERAL

The terminology used for dominating rock types in the Høydalsfjorden Complex reflects their metamorphic and deformational state. The rocks are affected by lower greenschist facies metamorphism (chlorite grade). This means, for example, that the *sedimentary* rock type "arenite" will appear as the *metamorphic* counterpart "psammite" in the study area. In addition, the rocks are also characterised by strong deformation. In the present work, the *deformation* will be indicated by the use of the prefix "*meta*-". Accordingly, the rock types will be designated "meta-psammite", "meta-pelite", etc. Regarding the sub-ordinate rock types, the term "phyllite" is used instead of "schist", for mica-rich rocks that has a "phyllitic" appearance; and the term "organic-rich shale" is used instead of "organic-rich schist", since the rock tend to have a "shaly" appearance in the field; and carbonates are termed marbles.

In the study-area, the meta-*sedimentary* Høydalsfjorden Complex (HC) rocks, which mainly consist of interbedded meta-greywackes/meta-semipelites; meta-psammitic to meta-semipelitic rocks; meta-psammites; etc., constantly occur in gradual transitional stages into each other. This makes mapping of the rocks very difficult, and on the main map (**Plate 1**) the rocks are therefore mostly grouped together. In certain areas, however, there appears to be one clearly dominating rock type, and such areas are indicated on the map (**Plate 1**). The study-area also contains minor amounts of other rock types (e.g. phyllites, meta-conglomerates, marbles and organic-rich shales), but these occurrences are too small to be displayed on the main map. They are therefore indicated with a letter on this map (**Plate 1**).

On the scale of **a few hundred metres**, the meta-sedimentary rocks in the HC are generally characterised by a very monotonous overall appearance, although the rock may gradually change character from place to place. The different rock types do not form distinct layers that can be mapped, and there is no lithologic or structural marker horizon in the complex. Combined with the lack of way-up control, this means that it is impossible to determine the mutual spatial positions of the rock types, and stratigraphic relationships between the rock types cannot be inferred.

The Høydalsfjorden Complex has been divided into 8 structural sub-areas, which are shown on the index map (**Plate 3**). The sub-areas have mostly been defined on the basis of naturally sub-dividing landscape-/topographical features like fjords and spurs of Devonian rocks. This subdivision will primarily be used in the structural analyses, (Sect. 4.3), but will also be used to ease the reader's localisation of place names on the main map (**Plate 1**). The sub-areas are "named" according to the compass directions to their positions as seen from the central part of the study area. The 8 sub-areas are: (1) WNW, (2) N and NE, (3) E, (4) SE, (5) S and SW, (6) WSW, (7) W, and (8) "central" (= the Osen meta-psammite inlier). Underlined letters, e.g. "N and <u>NE</u>", "<u>S</u> and SW", means that the reader is referred to this particular area within the sub-area.

4.2.1.2 DESCRIPTION OF META-SEDIMENTARY ROCKS

Interbedded meta-greywacke and meta-semipelite ("qz-fsp-mica-schist")

The rock consisting of interlayered meta-greywacke and meta-semipelite (**S_0**-layering) is particularly present in the WSW, W, WNW and <u>N</u>-NE sub-areas (**Plate 1**, **Plate 3**). This "two-component-rock" generally has a medium grey colour on fresh surface (**Fig. 4.2**) (although further to the west-northwest it may turn greenish) and a lighter grey colour on weathered surface (**Fig. 4.3**). The rock is locally calcareous. The meta-greywacke layers have thicknesses ranging from **5 cm** (**Fig. 4.3**) to **70 cm**, but are typically **10–40 cm** (**Fig. 4.2**). The meta-semipelitic layers between the meta-greywacke beds are usually up to **10 cm** thick (**Fig. 4.2**). This rock type can be particularly well studied at the Novene area (sub-area: WSW) (**Fig. 4.3**); at Gravaneset (sub-area: W); and at the area between Osstrupen (**Fig. 4.2**) and a line Teigen–Pollen–Barlindbotn (sub-area: WNW). The meta-greywacke layers may be interpreted as "turbidite" horizons. Vague indications of grading within layers has been seen occationally, but apart from the **S_0**-layering, no conclusive sedimentary structures have been observed. Locally, the rock may change to a more homogeneous meta-psammite, meta-greywacke or meta-semipelite. The main minerals in the rock are quartz, chlorite, epidote, albite, white mica, carbonate and opaques, with or without pyrite. Biotite is generally absent (with a few exceptions, Sect. 4.4.1). The meta-psammitic variants contain more quartz, albite and white mica, and the more meta-semipelitic variants more chlorite and epidote. The amount of carbonate may vary from **0–20 %**. These meta-sedimentary rocks contain a large number of intrusive meta-gabbro bodies, which are described below (Sect. 4.2.2.1). An **S_1**-cleavage is parallel to the **S_0**-layering. Occationally, the rocks have a schistose character (qz-fsp-mica-schist), with a fabric that has partly obscured the original sedimentary layering. The schistose rock type is prominent on the mountainside that is located between the Devonian rocks of eastern part of Vikafjell, and the shore at Sunnarviken (sub-area: <u>N</u>-NE).

Undifferentiated meta-psammitic to meta-semipelitic rocks ("qz-fsp-mica-schists")

The undifferentiated meta-psammitic to meta-semipelitic rocks are present in the N-<u>NE</u>, E, SE, and S-SW sub-areas (**Plate 1**, **Plate 3**). The rocks usually change very gradually in composition from meta-psammitic (**Fig. 4.4**) to meta-semipelitic (**Fig. 4.5**), and is locally calcareous. On weathered surface, the meta-psammitic versions are light grey (**Fig. 4.4**), whereas the more meta-semipelitic varieties typically have a darker grey colour (**Fig. 4.5**). On fresh surface the colour is grey or greenish grey. Varieties with a more pronounced greenish colour is present in the E sub-area. Lenses of vein quartz are locally abundant (**Fig. 4.4** and **4.5**). The main minerals in the rocks are quartz, albite, chlorite, epidote, white mica, carbonate and opaques, with a very variable amount of carbonate. The meta-psammitic variants contain more quartz, albite and white mica, and the more meta-semipelitic variants more chlorite and epidote. Structurally, the rocks are characterised by the strong tectonic planar $\mathbf{S_1}$-fabric, and the longest dimension of the vein quartz-lenses are oriented parallel to $\mathbf{S_1}$. This $\mathbf{D_1}$-deformation has usually altered the rocks into qz-fsp-mica-schists (**Fig. 4.4** and **4.5**), where the character of the schistosity depends on the protolith and the degree of deformation. The original bedding is often obscured. In the N-NE, E and SE sub-areas, rocks with an (**$\mathbf{S_3}$**-)mylonitic to ultramylonitic top-to-the-west $\mathbf{D_3}$-fabric define zones that may be up to **some tens of metres** thick. Such zones are particularly present in the mountainside between the Devonian rocks of Vikanipa and the fjord at Sunnarviken. No meta-gabbro has been observed in these rocks, with exception of the S-SW sub-area, where a single body was observed at the river Høydalselven near Høydalsstølen. Also the schistose variants display a gradual change into more feldspato-quartzitic variants (**Fig. 4.4**) or micaceous variants (**Fig. 4.5**). The schists particularly constitute large parts of the N-NE sub-area between the Devonian rocks of the eastern part of Vikafjell and the fjord of Eikefjorden, a sub-area where the rocks also tend to be mylonitic (Sect. 4.3.3.1). The schists are also present north of the Devonian rocks of Leirvågfjellet (sub-area: N-NE).

Meta-psammite (qz-fsp-schist)

The meta-psammite has a whitish colour on fresh surface and whitish to light grey colour on weathered surface. The rock is quartzitic to arkosic in composition. Thin meta-semipelitic bands may occur (**Fig. 4.6**), but the rock may also be without such bands on outcrop scale. The bands are interpreted as primary $\mathbf{S_0}$-layering. The main minerals in the rock are quartz, albite, white mica, chlorite, with epidote and opaques as accessory minerals. The tectonic planar $\mathbf{S_1}$-fabric is parallel to $\mathbf{S_0}$, and in combination they create a strong composite $\mathbf{S_0}$-/$\mathbf{S_1}$-foliation. Major areas of meta-psammites are located several places within the study area: In the area between Pollen and Litlevika (sub-area: WNW), meta-psammitic rocks are present locally. Local occurrences are also found to the west of the Devonian rocks that constitute the summit of Mannen (sub-area WNW), where the meta-psammitic layers alternate with layers of meta-semipelite, and gives the impression of having been turbidite deposits (**Fig. 4.7**). The rocks are also locally present at, and to the west of, the Høgdene area (sub-area N-NE) (the latter rocks are not marked on **Plate 1**). Most parts of the Høgdene inlier (Sect. 4.2.1.4 and Sect. 5.4.2), which is exposed within the Devonian rocks on the northern mountainside of Vikafjell, (sub-area: N-NE), consists of meta-psammite, as does the Osen inlier (Sect. 4.2.1.4 and Sect. 5.3) which crops out within the central parts of the Devonian massif (sub-area: "central"). Between Teigafjell and Nonsnova, meta-psammitic rocks are present in the eastern slopes of the wedge-shaped outcrop defined by HC-rocks. The rock is also present at Osknoltren about **800 m** to the east-northeast of the summit of Vikanipa (sub-area: N-NE). At Fjellsenden (sub-aera: SE) (**Fig. 4.6**), the rock constitutes the whole ridge located between the Devonian deposits in the west and the river Kleiveelva to the east (**Plate 1**). The rock is also found to the south and west of the summit Litledokka (consisting of Devonian rocks) (sub-aera: SE). No sedimentary structures have been observed in the meta-psammites, apart from the lithological layering. In thin-section, the rock is completely recrystallised with no sign of a sedimentary texture. With incresing degree of deformation, the rock turns into a qtz-fsp-schist.

Reconnaissance investigations to the west of the study area have confirmed that several large occurrences of meta-psammites are also located in these areas (see e.g. the geological map of Kildal 1970). On the islands of Stavøya and Alvora (**Fig. 2.4**), the meta-psammites show gradual primary transitions into surrounding meta-greywackes and meta-semipelites of the same type as in the study area, and the rocks are generally strongly deformed. The Sandvika meta-psammite on the peninsula between Eikefjorden and Høydalsfjorden (**Fig. 2.4**) is locally mylonitised, but in spite of this, primary sand grains and gritty grains may be observed at the pier in Sandvika (**Fig. 4.8**).

Greenschist

Greenschists are present in the southwestern part of the study area, notably on the peninsula of Langeneset and further eastwards (sub-area S and SW) (**Plate 1**), as also seen on the geological 1:250.000 map Måløy (Kildal 1970) and the 1:50.000 map Eikefjord (Bryhni & Lutro 1991a, 2000a). Small occurrences are also found at Kalsvik (sub-aera: WNW) (**Fig. 4.9**), and presumably also at the Litleteigen inlier (sub-area: E) at the southern part of the lake Vassetvatnet. The best exposures are present on the Langeneset peninsula. Apart from the strong green colour, the greenschist at Langeneset resembles the meta-greywackes, meta-semipelites and meta-psammites that are present elsewhere in the Høydalsfjorden Complex, and the rocks thus appear to be of essentially meta-sedimentary ("turbiditic") origin with a profound basic volcaniclastic component; although a pure volcanic origin is possible for parts of the rocks. The main minerals in the rock are chlorite, epidote, actinolite, white mica, quartz, albite, and opaques, with or without carbonate and pyrite. The more light green "felsic" variants contain more quartz, albite and white mica, and the dark green "mafic" variants more chlorite, epidote and actinolite. In the Solund–Stavfjorden area, Furnes et al. (1990) interpreted similar greenish rocks as basic volcaniclastic sequences. In the study area of the present thesis, the rocks may be termed "meta-greenwacke". The green colour is caused by a high content of chlorite, epidote and actinolite. At the Langeneset peninsula, epidote nodules are present in the rock. In addition, possible "felsic dykes" are present (Sect. 4.2.2.3).

Phyllite

Phyllite is marked with a "p." on the main map (**Plate 1**). However, the rock is not very abundant in the study-area. Phyllite may be studied in road sections along the road that departs from county road no. 542 (Fylkesvei 542) and leads to the timber shipment pier located **a few km** to the west of Sunnarviken. The rock is also locally present on the hillside between the Devonian rocks of Vikafjell and the shore of Eikefjorden (sub-area: N-NE), but these occurrences are not indicated on the main map (**Plate 1**).

Conglomerate

Meta-conglomerates have been observed in the study area, and are marked with a "c." on the main map (**Plate 1**). Since the rocks are generally little deformed, they will here be termed "conglomerates" without the "meta-" prefix. As discussed below, variable degrees of uncertainty may exist as to whether the conglomerates are part of the Høydalsfjorden Complex (HC) or the Håsteinen Devonian Massif (HDM).

Conglomerates that with a fairly high degree of *certainty* belong to the HC can be termed "***Fairly certain HC-conglomerates***" (see below), and such conglomerates have been observed at two localities (**Plate 1**); (1) to the northeast of the Devonian rocks of Vikanipa (sub-area N-NE), and (2) to the west-southwest of the Vasset Farm (sub-area: N-NE). Moreover, conglomerates that *possibly* belong to the HC (termed "*Possible HC-conglomerates*", below) have also been observed at two localities (**Plate 1**): (1) to the north of the Stigen Devonian spur (sub-area: N-NE), and (2) to the north of the Vikanipa Devonian spur, where the possible Høydalsfjorden Complex conglomerate is located only **a few metres** from proper Devonian conglomerate. The conglomerates will be desribed in the following.

***Fairly certain HC-conglomerates*:** The first locality containing fairly "certain" HC-conglomerates is present on the mountainside about **500 m** (horisontal map distance) to the northeast of the summit of Vikanipa. The HC conglomerate forms a steeply dipping cliff-like exposure which is up to **6 m** high and **20 m** long, and which is surrounded by recently formed talus on all sides, so that any potential contacts to surrounding rocks are covered. The nearest exposure of a surrounding rock is a qz-fsp-mica-schist located about **5 m** northwest of the outcrop. The conglomerate is only slightly deformed. A weak, but clear, tectonic cleavage goes through the rock, and this cleavage occationally bends around the clasts. Longest dimension of the clasts tend to be oriented parallel to the cleavage (**Fig. 4.10**), and contributes to the mesoscopic tectonic fabric. The cleavage has a strike/dip orientation of about **092/58 S**, which is roughly parallel to the cleavage of the ordinary HC rocks in the area. The conglomerate is clast supported, poorly sorted, unbedded, without grading or other sedimentary structures, and polymict, with the clasts consisting of about **50 %** grey-coloured qz-fsp-mica-schist; **30 %** whitish meta-psammite (of both foliated and more massive and coarse grained types); **10 %** vein quartz; and in addition a few clasts of dark greenschist. Typical clast sizes range from **1 x 1** to **6 x 3 cm**, and the maximum size observed was **25 x 18 cm**. The meta-psammitic clasts have the largest diametres. The rock has a continuous range of clast sizes, and although it is locally slightly more "sandy" between the pebbles, the conglomerate has no bimodal clast size distribution. Most clasts are angular, but the larger clasts tend to be slightly better rounded. Clasts with pre-depositional internal foliations tend to have elongated shapes, whereas less foliated clasts are less elongated. The "matrix" is carbonate-bearing, displaying an orange colour on weathered surface and a dark grey colour on fresh surface.

To some degree, the conglomerate resembles the Devonian conglomerate, and could theoretically be of Devonian age. However, to be a Devonian conglomerate, the rock would most likely would have to be a downfallen block coming from the Vikanipa Devonian spur. The exposure is, however, situated too far east to be "en route" for blocks falling from the Vikanipa Devonian spur, and it is therefore not likely that it is a downfallen block. In addition, all the surrounding talus blocks are qz-mica-schists, and none are conglomeratic, suggesting that blocks from Vikanipa would be totally exotic in this area. More importantly, the presence of a penetrative cleavage also makes this rock different from the proper Devonian conglomerates. All considered, the conglomerate may therefore be interpreted as an integral part of the substrate.

The second locality displaying fairly certain HC-conglomerates is present on the hill-slope **350 m** (horisontal map distance) to the southwest of the Vasset Farm (**Plate 1**) (sub-area: N-NE). The exposure is very limited, but a brief description will be given. The conglomerate is clast supported, unbedded, and polymict. The dominating clasts are foliated and massive meta-psammite and vein quartz. Due to lack of exposure, the number of pebbles observed are limited. The clasts are angular to subangular, and foliated clasts are elongated. The conglomerate has a continuous range of clast sizes, with clasts having diametres **0–10 cm** and with **5 cm** as an average, and is not bimodal. The finest grain fraction may nevertheless be considered as a matrix, and this matrix is blackish to orange on weathered surface and dark grey to black on fresh surface. No sedimentary structures can be observed. The conglomerate is surrounded by greenish meta-psammites/quartz-schists, but contacts to these rocks are not exposed. A weak cleavage with orientation **096/43 S** is developed in the conglomerate, and the longest dimension of the clasts are usually oriented parallel to this cleavage, with the cleavage wrapping around some clasts. The cleavage is roughly parallel to the cleavage in the HC rocks of the area, and the presence of this cleavage suggests that the conglomerate belongs to the HC.

***Possible HC-conglomerates*:** The first to be discussed of the two conglomerates that *possibly* belong to the HC, is a conglomerate present on the northern side of the Stigen Devonian Spur (subarea: N-NE) (**Plate 1**). Befor proceeding to the “possible HC conglomerate” itself, it is worth noting that the conglomerate of this spur is a *proper* Devonian conglomerate that form a roughly W-E trending and east-sloping topographic ridge with a northern escarpment that inclines steeply to the contact against the subjacent HC rocks. Generally in the Håsteinen area, the Devonian rocks often form this kind of steep “contact escarpments” above the HC rocks. The *possible* HC-conglomerate discussed here has its southern “termination" situated just **a few metres** to the north of the "contact escarpment" formed by Devonian rocks. The eastern boundary of the exposure of the possible HC-conglomerate is positioned roughly **50 m** (horizontal map distance) to the west of the road; and in the hillside, the conglomerate crops out at altitudes ranging from **c. 50 to 100 m**. The **few-metres-wide** strip of ground that separates the conglomerate and the Stigen Devonian spur contains no rock exposures. The possible HC conglomerate is generally located on a quite steep, wooded east-dipping slope; which makes mapping difficult. Scattered exposures of this conglomerate are mainly present within an area of about **30 x 30 m** as measured on the steep slope surface itself. No exposures of contacts to the surrounding rocks have been found. The clast sizes are usually **5–10 cm** in diametre, with the largest clast being **10–20 cm** across, but a continuous range of clast sizes, from the smallest to the largest size is present, i.e. the conglomerate is not bimodal. The conglomerate is clast supported, poorly sorted, apparently unbedded, and polymict with clasts of qz-fsp-(mica)-schists, foliated and massive meta-psammite, and vein quartz. The clasts are mainly sub-rounded, with some clasts being sub-angular and rounded. The conglomerate appears to be ungraded and without any sedimentary structures. Sand layers have not been observed. The clast shapes range from near "spherical" to a clast axis ratio of **1:2**. No cleavage appears to be present within the conglomerate.

The conglomerate differs from the nearby situated Devonian conglomerate, by a much smaller clast size. In addition, the conglomerate is located to the north of, and below, the steep "contact escarpment" formed by the Devonian rocks proper. Normally, the areas located below such “contact escarpments” contain rocks of the HC. Based on these features, it is here tentatively concluded that the conglomerate is part of the HC.

However, due to the lack of deformation, the conglomerate also resembles the Devonian conglomerates present in the near by Stigen Devonian spur. Since proper Devonian conglomerate is present so near, it has not been possible to determine with certainty that the conglomerate really belongs to the HC.

The second conglomerate which theoretically *could* belong to the HC, is the conglomerate that forms the eastward-directed little spur situated in the north-dipping mountainside just to the north of the summit of Vikanipa (sub area: N-NE). Although the little spur has been interpreted as Devonian on the main map (**Plate 1**), some degree of uncertainty prevails conserning the interpretation, justifying the following discussion. The spur-conglomerate is exposed in a **c. 50 m** long and **3–6 m** high, vertical cliff striking essentially east–west. Lichen covers parts of the outcrop. The conglomerate is separated from the main mass of Devonian conglomerates by a poorly exposed zone containing mylonitic rocks interpreted as part of the HC. The distance between the spur-conglomerate considered here and the main mass of Devonian rocks is **50 m** in the eastern parts of the spur, being gradually reduced to about **10–20 m** in the western parts. The poorly exposed “separation-zone” forms a flatter "shelf" on the otherwise steep hillside, and a potential contact between the two conglomerates is not exposed. The conglomerate in the little spur may therefore theoretically belong to the HC. The conglomerate is clast supported, very poorly sorted, unbedded, wihtout sedimentry structures, and polymict. Foliated and massive meta-psammite constitute more than **80–90 %** of the clasts. The conglomerate contains a continuous range of clast sizes from pebbles, via cobbles to boulders. The clasts are usually sub-angular, but angular and rounded clasts are also present. Foliated clasts have an elongated shape with an axes-ratio up to **1:2**, whereas massive/unfoliated clasts are more equidimensional. The conglomerate is very similar to the adjacent main mass of Devonian conglomerate which belongs to the Vikafjell Formation (Sect. 5.2.2.1) of the Håsteinen Devonian Massif. At the northeastern edge of the spur, a vague foliation is present in the conglomerate. It is difficult to decide whether this foliation is the pre-Devonian major HC-foliation (**S_1**), which would make the conglomerate itself *pre-Devonian* and part of the HC, or whether the foliation is related to shear that post-daties formation of the Devonian rocks, possibly making the conglomerate Devonian. Apart from the vague foliation at the northeastern edge, the conglomerate appears to be completely undeformed. Hence, because of its strong similarity with the Devonian conglomerates, its closeness to the conclusive Devonian rocks, its large size, and the apparent absence of penetrative deformation, this conglomerate is here interpreted as belonging to the Håsteinen Devonian Massif, although it cannot be completely excluded that it belongs to the HC.

Marble

Only very minor amounts of marble are present in the HC within the study area. Marbles have been observed at several places, and are marked with an "m." on the main map (**Plate 1**). Massive marble have been observed at ten localities: (i) At Osskåra to the east of Vikanipa (sub-area: N-NE), the marbles form lenses with orange weathering colour. The lenses are typically up to **5 cm** thick and **50 cm** long and have an orange weathering colour (**Fig. 4.11**). Marble are also found on the immediate northeastern side of the Stigen Devonian spur (sub-area: N-NE). (ii) Here, to the east of the roadcut for instance, marble with ash-grey weathering colour is present over a distance of **5 m** along the lake shore. (iii) Further west, i.e. up on the hillside along the Devonian spur, lenses with orange weathering colour are present. The lenses are up to **20 cm** thick, and may be **2**

m long. (iv) A marble lens is also present at one locality at the southern side of the Stigen Devonian spur (sub-area: E). (v) On the immediate northern side of the Strupeneset Devonian spur (sub-area: E) at the southern part of lake Vassetvatnet, lenses of marble with grey weathering colour are present at the lake shore to the east of the road. (vi) Lenses are also present several places high up along the "Devonian/substrate contact-cliff ". (vii) In the western parts of the Høgdene meta-psammite inlier (Sect. 4.2.1.4), marble-"bodies" with a size of about **1 m** thickness and unknown length are present. These lens-like bodies have an orange weathering colour. (viii) The Osen meta-psammite inlier (Sect. 4.2.1.4) contains marbles with grey-white weathering colour at the northwestern-most part of the "eastern outcrop" of the Osen meta-psammite (i.e. the outcrop east of the Osen farm), at a locality **20 m** to the east of the tractor road. (ix) Marbles are also present a few other places along the northern side of the Osen inlier. (x) In addition, high densities of much smaller marble lenses are locally quite abundant in the meta-greywacke/meta-semipelite area in the WSW and WNW sub-areas, having a white colour on fresh surface (**Fig. 4.12**). These lenses appear to be distributed in high densities throughout larger volumes of rocks, instead of being concentrated as few descrete “lense-bodies” in certrain zones.

Organic-rich shale

Organic-rich shales have been observed at four places in the thesis area, and are marked with an "o." on the main map (**Plate 1**). The rocks have a black colour, and split into thin flakes along the tectonic cleavage. The shale occur (i) in small outcrops at the road section present along the western side of the lake Vassetvatnet in the interval **100–200 m** north of the Stigen Devonian spur (**Plate 1**) (sub-area: N-NE). The rock may also be observed (ii) in the road section to the east of the village of Svardal (sub-area: SE), as well as (iii) on a flat "shelf" in the hillside between the summit of Vikanipa and the shore of Sunnarviken (sub-area: N-NE). The shale is also present (iv) on the steep slope just to the northeast of the meadows of the summer farm of Vassetstølen (sub-area: N-NE). The outcrops appear to be isolated and do *not* form a continuous marker horizon.

4.2.1.3 “DISTURBED LAYERING” (S0), AND “DISRUPTED ROCKS”

“Disturbed layering”.

(Probably caused by soft-sediment slumping, “D_0-deformation”)

In the western sub-areas containing the turbidite-appearing interlayering of meta-greywacke and meta-semipelite, a strange form of disturbance has effected the layering. The otherwise continuous layering may be "broken up" and reoriented in the most peculiar ways (**Fig. 4.13**). These rock features are possibly a result of syn-sedimentary (**D_0-**) slumping and sliding processes, or possibly related to intrusion of numerous gabbro-bodies at the pre-collisional oceanic stage (see Sect. 4.2.2.1). However, conclusive *evidence* for soft-sedimentary deformation is difficult to find due to the later overprinting of orogenic deformation and lack of continuous exposures (see Sect. 4.3.3)

"Disrupted rocks".

(Probably caused by D_2-deformation)

The rocks which are here grouped as "disrupted rocks" are characterised by a "fragmented" appearance where the main S_1-foliation is less obvious and the rock appears to contain a lot of "fragments" that resembles pieces of breccia clasts. These "fragments" are typically **5–15 cm** across, but may range up to a size of **10 x 40 cm**. The shapes are both equidimensional and elongated. The fragments usually have an internal S_1-foliation and give the impression of having been an integral part of the S_1-foliated rock prior to the "disruption". The S_1-cleaved meta-greywacke layers frequently terminate abruptly at high angle towards other layers, both in fold-appearing structures (**Fig. 4.14**) and elsewhere. Smaller "fragments" of S_1-foliated meta-psammite are rotated as compared to the rest of the rock fabric (**Fig. 4.15**). Parts of the rock may have preserved the continuity of the original S_1-foliation, whereas adjacent parts may be completely disrupted (**Fig. 4.16**). The disrupted features occur only locally, and have particularly been observed in the area between the shores of Høydalsfjorden and Eikefjorden (sub-area: WNW), and further eastwards to the area between Høgdene and the shore of Eikefjorden, as well as in the Osknoltren area (sub-area: N-NE).

Discussion/interpretation: It is not a straight foreward matter to interpret these features. The possibility that some sort of brecciation process related to brittle faulting should be responsible for the ”disrupted” rocks, is not likely since the disruption does not follow spesific zones. It might be suggested that the rocks were originally a sedimentary breccia with foliated clasts that has later been deformed along with the rest of the rocks. But this explanation does not seem appropriate either, as it is the S_1-fabric which has been disrupted. It is possible that during (F_2-) folding (Sect. 4.3.5.2), the competence contrast between the meta-greywacke and the meta-semipelite have resulted in "disruption" of the competent meta-greywacke layers, and indications of this are seen in some folds. Locally, the structures may, however, appear to be too "chaotic" to allow competence contrasts to be the only explanation, particularly since more homogeneous meta-psammites, appearantly without any particular competence contrasts, are also "disrupted" (e.g. at Osknoltren, sub-area: N-NE). It is also possible that some of the "fragment"-structures are due to regolith formation at the original sub-Devonian palaeo-surface. The disruptions could not have been caused by violent processes related to the numerous intrusive meta-gabbro bodies in the meta-sedimentary rocks in the area, since the gabbros were intruded prior to the formation of the S_1-foliation (see Sect. 4.2.2.1). In addition, it could be suggested that the D_3-shear that formed the mylonite zones between Vikafjell and Høydalsfjorden, played a role in the “disruption”, but this is not obvious since the disruption does not seem to occurre along these zones.

Of the above mentioned possibilities, it is considered most likely that a large portion of the disruption has been caused by competence contrasts during the (post-D_1) F_2-folding. The relativly low metamorphic grade of the rocks, i.e. lower greenschist-facies (Sect. 4.4), implies that the rocks were not very far from the brittle-ductile transition, implying P/T conditions which could have facilitated brittle behaviour, particularly of feldspar-rich meta-psammites. Other factors, as for example elevated fluid pressures, may have contributed to the disruption process, but it has not been possible to prove or disprove this. Late brittle faulting may be excluded as a contributing process.

4.2.1.4 INTRA-DEVONIAN INLIERS

Three inliers of HC-rocks are present within the Håsteinen Devonian Massif (**Plate 1**). These are (i) the Høgdene meta-psammite inlier along the northern margin of the HDM, (ii) the Osen meta-psammite inlier in central parts of the HDM, and (iii) the Litleteigen greenschist inlier near the southern end of lake Vassetvatnet. The two first inliers are completely surrounded by Devonian rocks. The last one, in addition to being surrounded by Devonian rocks, is also partly bounded by the lake Vassetvatnet, but the outcrop is here nevertheless treated as an inlier. Contact relations to the Devonian rocks, as well as structural models and interpretations of these outcrops are given in Sect. 5.4. Only rock descriptions will be presented in the following.

Høgdene meta-psammite inlier

The Høgdene meta-psammite inlier is located in the steeply northward-dipping mountainside close to the northern margin of the HDM (**Plate 1**). The outcrop is about **1.3 km** long in the W-E direction, up to **100 m** thick as measured on the montain side, and has a structural thickness of up to **90 m**. The steep parts of the outcrop are inaccessible, and the western part gives the easiest access. The meta-psammite has an arkosic to quartzitic composition. The colour is whitish on fresh surface, and greyish white on weathered surface (**Fig. 4.17a**). Structurally, the rock is dominated by the $\mathbf{S_0}$-parallel $\mathbf{S_1}$-foliation (**Fig. 4.17a**), although more massive/unfoliated parts are also present. Lenses and more continuous bands of vein-quartz lie within the foliation (**Fig. 4.17b**). The $\mathbf{S_1}$-foliation is locally folded by $\mathbf{F_2}$-folds (**Fig. 4.17b**). In the areas just to the east of "Galmannsskåra", minor areas of meta-semipelites as well as marbles (**Fig. 4.18**) are present. Locally, the meta-psammite contains a breccia structure that is confined to a limited area and therefore not related to faulting (**Fig. 4.19**). This breccia is a "proper" breccia, and does not resemble the «disrupted rocks» described in Sect. 4.2.1.3. The breccia has been observed to change gradually into unbrecciated rock. The breccia is interpreted as a Devonian regolith formed near to the sub-Devonian palaeo-surface.

The regolith interpretation implies that the breccia formed *in situ* on the original sub-aerial rocks surface, from some sort of weathering or other decay process affecting the rock at the place (e.g. Kearey 1996). In the present thesis, the regolith interpretation is taken to imply that the breccia experienced none or only minimal transport, and that it remained unconsolidated until it was burried by “proper” Devonian sediments.

Osen meta-psammite inlier

The Osen meta-psammite inlier is located between Osen in the west and the summit of Teigafjell in the east. The extensive Quaternary deposits at Osen divides the meta-psammite inlier into two parts, a major "eastern outcrop" and a minor "western outcrop" (**Plate 1**). East of the Devonian rocks that form the mountain of

Teigafjell, a wedge-shaped outcrop of meta-psammite and meta-"semi"-psammite reaches, from the east, "up into" the Teigafjell Devonian rock area, and this wedge is interpreted as the continuation of the Osen inlier (**Plate 1**). The Osen meta-psammite has a whitish colour on fresh surface and whitish to light grey colour on weathered surface (**Fig. 4.20**). The composition is arkosic to quartzitic. Other rock types than meta-psammite have, with exception of scattered carbonate slivers along the northern margin, only been found near the Osen farm, at the northwesternmost corner of the "eastern outcrop". Here, a small area contains rocks with a more grey-green meta-semipelitic component (**Fig. 4.21**), as well as very small occurrences of marbles. This somewhat more meta-semipelitic rock is very similar to the HC-rocks located in the mountainside of the N-NE sub-area.

Structurally, the rock is dominated by the $\mathbf{S_1}$-foliation which is folded by $\mathbf{F_2}$-folds (**Fig. 4.20**, **4.21**). The meta-psammite is locally brecciated (**Fig. 4.22**), with the breccia occurring within irregular “spots ” that measure up to a **few square metres** in size within the rock surface. The breccia does not form zones that could be related to brittle faulting. When contacts to the surrounding meta-psammitic rocks are exposed, it may be observed that the breccia changes gradually into the unbrecciated rock. The breccia is interpreted as related to the development of a sub-Devonian regolith. At Teigafjellet, the substrate-wedge that outcrop-wise “comes up” from the east into the Teigafjellet mountain plateau, is meta-semipelitic to meta-psammitic in the upper parts, and clearly meta-psammitic further down the eastern mountainside. The wedge is also locally brecciated. These features show the clear similarity between the substrate wedge on Teigafjellet and the Osen meta-psammite inlier further west, and justifies the conclusion that they are connected below the cover of Devonian sediments on Teigafjellet.

Litleteigen greenschist inlier

The Litleteigen greenschist is located at the southernmost end of lake Vassetvatnet, in the mountainside to the west of the lake. An eastward-dipping recent talus-cone rises steeply from the lake shore and up the mountainside. Along the southern limit of this cone, an E-W oriented vertical cliff wall is present. The base of this cliff wall is defined by the surface of the talus cone which dips steeply to the east. The inlier rocks are present within this vertical cliff wall, and may be studied along the steeply dipping base of the cliff wall, adjacent to the upper area of the talus cone. Access to the outcrop is difficult. The small outcrop consists of a dark-greenish serpentinite/greenschist appearing rock. Further up into the vertical cliff, a greyish greywacke-type rock appears to be present. Vertical white stripes on the cliff wall, presumably formed from running water, may indicate that carbonates are present further up in the vertical cliff.

4.2.2 META-IGNEOUS ROCKS

Meta-gabbro is the main igneous rock in the studied part of the HC. In addition to this rock type, "felsic layers" that may be interpreted as dykes, are present. Possible pillow lavas are located near the sea northwest of the Osstrupen road-bridge/Straumsnes area. A syenite body is located along the southern margin of the Håsteinen Devonian Massif. Quartz-epidote veins are also present in the study area, but these veins “intruded” after deposition of the Håsteinen Devonian Massif and are treated in Sect. 5.2.3.

4.2.2.1 META-GABBRO

Description

Bodies of meta-gabbro have only been found in the WSW, W, WNW and N-NE sub-areas (**Plate 1, Plate 3**). In addition to this, *one* body was observed in the S-SW sub-area in the river Høydalselven just to the east of Høydalsstølen. On the main map (**Plate 1)**, the presice location of two examples of such bodies are indicated with the letter "g."; notably at the Gravaneset headland (sub-area: W) where the body is marked by a single ”g.”, and **500 m** to the north of the Osstrupen road-bridge (sub-area: WNW), where the body is marked by three “g.”s. The body situated at the very tip of Gravaneset has, in addition, been mapped in detail on the 1:1000 map (**Plate 4**), together with examples of several other bodies in the same area. Also the 1:200 map (**Plate 5**), being an enlargement of parts of the 1:1000 map (**Plate 4**), shows the detailed outline of a few meta-gabbro bodies. Although grain-sizes in these rocks may occationally be almost basaltic rather than gabbroic, the rocks will for simplicity be referred to as meta-gabbros in the following.

The meta-gabbroic rocks are present as apparently isolated bodies that are surrounded by the meta-sediments of the HC. The size of the bodies may be from **a few metres** to at least a **few hundred metres** across, but they are typically **some tens of metres** in diametre. The bodies are present in a great number, and their size is generally too small to allow mapping, with the limited degree of exposures present. The distance between the bodies may vary from **a few metres to hundreds of metres**. The shape of the bodies are generally very difficult to observe due to the lack of exposures. Mesoscopically, the rocks usually have a typical "gabbroic" appearance, but the rock may also be a micro-gabbro, a basalt or a gabbro-pegmatite (**Fig. 4.23**). The main minerals in the rock are retrograded hornblende, actinolite, cerisitized feldspar, chlortite, epidote, zoisite, white mica and opaques. Biotite is generally absent, but was observed at one occation.

Xenoliths of meta-semipelite and meta-greywacke are present in the meta-gabbro bodies. Small xenoliths may be partly melted (**Fig. 4.24**, **4.25**). Otherwise, xenoliths have a lithologic banding, but do not appear to have a profound tectonic foliation of the same type as the meta-sediments of the wall rocks. This indicates that the meta-gabbros intruded at an early stage (see below). The meta-gabbro may be either massive (**Fig. 4.23**), or partly to completely penetrated by the **S_1**-foliation (**Fig. 4.26**).

The contacts towards the meta-sediments are usually preserved as primary intrusive contacts. The bodies frequently show grain size reduction towards the margin which must be interpreted as chilled margins (**Fig. 4.27**). Tectonic modifications of these contacts by shear are common.

The wall rock is usually a meta-semipelite or a meta-greywacke. Contact metamorphic aureoles are developed in the meta-sediments bordering the meta-gabbro bodies, but the aureoles do not appear to exceed a thickness of **15–20 cm** as judged in the field. Some melting of the wall rocks and back-veining into the meta-gabbros may be present (**Fig. 4.28**). It has been observed that the intrusive contacts of the meta-gabbros cut the **S_0**-layering of the meta-sediments, and that both rocks have later been overprinted by the S_1-fabric (**Fig. 4.29**).

Time and place of intrusion: discussion/conclusion

The gabbro bodies appear to have gone through the same phases of deformation as the meta-sediments, since the **S_1**-foliation is developed across the bodies, and since the bodies are never observed to clearly cut the tectonic **S_1**-fabric in the meta-sediments.

The meta-gabbros therefore probably intruded the turbiditic meta-sediments during the oceanic, pre-collisional stage. This may have occurred even while the sediments were still "wet" and unconsolidated. Examples of recent basic rocks that have intruded into wet sediments near oceanic spreading ridges are known from e.g. Bay of California (Einsele 1982, 1985). If the meta-gabbros intruded "wet" sediments, the release of water as explosive steam may have contributed to the soft-sedimentary "disturbance" (not disruption!) of the sedimentary layering as described above (Sect. 4.2.1.3). The meta-gabbros can be interpreted as a result of an "ophiolite" intruding its own primary cover of turbiditic greywackes and semipelites during the oceanic stage.

This, however, does not prove that the Høydalsfjorden Complex as a whole is part of the cover sequence to the Solund–Stavfjorden Ophiolite Complex. Since large areas of the Høydalsfjorden Complex are without meta-igneous material, it is possible that these areas are part of a melange. This is further discussed below (Sect. 4.2.3).

4.2.2.2 POSSIBLE PILLOW LAVAS

Pillow-like structures are present at the north end of the very narrow inlet that extends northwards from the fjord of Høydalsfjorden, at a position located about **500 m** to the north-northwest of the Osstrupen road-bridge (sub-area: WNW). A number of **7 or 8** pillow-like bodies consisting of meta-basalt are found, all of them confined within an area measuring **1 x 1 m**. This area is part of a larger outcrop of exposed rock measuring **several tens of square metres**. The "pillows" are elongated, with their longest dimension typically **40–70 cm** and a thickness of **10–20 cm** (**Fig. 4.30** and **4.31**). They are separated by **5–10 cm** thick layers of meta-sediments. These examples are the only pillow-like structures which have been observed during the field work.

4.2.2.3 POSSIBLE FELSIC DYKES

Possible felsic dykes have been observed at two areas; (i) in the greenschists at Langeneset peninsula, and (ii) at the peninsula of Holmesund which is located on the northern side of Høydalsfjorden, about

5 km to the west-northwest of the road-bridge crossing Osstrupen (**Fig. 1.1**). The "dykes" are usually parallel to the major **S_1**-foliation, and have been deformed by the **D_1**-deformation. However, the "dykes" have not been investigated in the present work.

4.2.2.4 THE KVANGAGJELET SYENITE

The Kvangagjelet syenite is located along the southern margin of the Håsteinen Devonian Massif (**Plate 1**), where it is very well exposed. The name of the syenite is taken from the river-canyon called Kvangagjelet, which follows the southern margin of the syenite body at its western part. Structural models for the localisation of this syenite will be discussed in Sect. 5.4.5. In the following, only a short description of the rock will be given. The body has not been radiometrically dated. Although the composition of this magmatic rock may vary from syenittic to monzonittic (Bryhni & Lutro 1991a, 2000a), it is mainly a syenite (Kolderup 1925, 1928), and the body will hereafter thus be termed a syenite.

The body forms an outcrop which is at least **1.5 km** long in the NW-SE direction, **180 m** wide in map plane, and **250 m** thick when measured on the mountainside of the Kvangagjelet canyon. The northern margin of the body is defined by the contact against the Håsteinen Devonian conglomerates, and the conglomerates are situated unconformably on top of the syenite. The southern margin is defined by a fault, along which the canyon of Kvangagjelet has been formed. This fault separates the syenite to the north from the qz-fsp-mica-schists of the HC to the south. The presence of this fault, together with an insufficient degree of exposure along the same fault, make the relationship between the syenite and the HC-schists uncertain.

The syenite has a whitish colour on weathered surface and a light pink colour on fresh surface. The pink colour is due to the high content of K-feldspar. The main minerals in the syenite are K-feldspar (microcline), plagioclase (partly to completely cerisitized), chlorite, epidote, quartz, white mica, rutile and opaques. Thin-section studies of a sample from the upper part of the river flowing from lake Håsteinsvatnet, showed that the original igneous texture may be quite well preserved, with only some degree of cerisitisation and chlorite-growth to indicate alteration. The grain size at this locality is **1–2 mm** for most grains, and the largest grain observed was less than **c. 3 mm**. Locally, the minerals in the rock have suffered extensive polygonisation, and numerous small subgrains then tend to completely modify the original igneous texture. In such instances, extensive growth of chlorite, epidote and cerisite have taken place. The rock is occasionally cut by ductile shear zones. The syenite is generally massive and unfoliated, but towards the southern margin the rock takes on a weak foliation orinted with a strike going WNW-ESE to NW-SE and a dip of about **60° NNE to NE**. The massive/unfoliated parts of the rock show signs of brecciation, and towards the southern marginal fault, this brecciation appears to increase. The brecciation is possibly related to the movements on the fault along the southern margin of the syenite. However, areas with brecciation may also occur in other parts of the body, for example in the southeastern parts. It cannot be excluded that parts of the brecciation was related to shock effects that could have appeared if the body was a land-slide into the Håsteinen Devonian Basin at the time of sediment deposition. Clearly intrusive veins or apophyses going from the body and into the HC qz-fsp-mica-schists to the south of the body, have not been observed. Kolderup (1925) made a brief field description of the rock, and stated that it resembled the syenites that forms the substrate to the Kvamshesten Devonian deposits. The syenites there are presently assigned to the Precambrian Dalsfjord Suite (/Complex), which Corfu and Andersen (2002) dated to **1634 +/-3 Ma**, interpreted as the age of intrusion.

In summary, no unequivocal evidence, that the Kvangagjelet syenite is part of the HC, has been found. If the syenite should be part of the HC, it would have to be either intrusive into the surrounding meta-sediments, or a mega-olistolith in a melange (see Sect. 4.2.3 below). Likewise, no field relations indicate that the body is a 'basement window' where the Precambrian rocks below the Høydalsfjorden Complex (HC) appear at the daylight surface. Rather, since the Precambrian rocks are strongly mylonitic both to the north and south of the HC (part of the NSDZ), and since the Kvangagjelet syenite is in fact largely unfoliated, the syenite body is most likely not a 'basement window'. On the other hand, it is possible that the body is a mega-sized land-slide block within the Håsteinen Devonian Massif. This is discussed in Sect. 5.4.5.

4.2.3 INTERPRETATION OF DEPOSITIONAL ENVIRONMENT FOR THE HC-ROCKS

Rock types (summary)

The meta-sedimentary rocks of the Høydalsfjorden Complex are dominated by (i) grey- or green-coloured sequences of interbedded layers of meta-greywacke ("turbiditic") and meta-semipelite, which are particularly present in the WSW, W, WNW and N-NE sub-areas**;** and (ii) undifferentiated meta-psammitic to meta-semipelitic rocks which are present elsewhere within the study-area (N-NE, E, SE, S-SW and "central" sub-areas). Light- or whitish-coloured homogeneous meta-psammite is present particularly in the "central" and SE sub-areas. In addition, minor occurrences of phyllites, conglomrates, marbles and organic-rich shales are present locally, mainly in the N-NE and E sub-areas. Greenschists (probably meta-volcaniclastics) are present particularly in the S-SW sub-area. "Felsic layers" in these greenschists are possibly dykes, but have not been studied in the present work. The meta-sediments are locally calcareous. Pillow lava have generally not been observed, with exception of one exposure measuring **1 x 1 meter**, containing a group of **7–8** pillow-like "bodies" in the WNW sub-area. Intrusive bodies of meta-gabbro are present in the meta-sedimentary rocks in the WSW, W, WNW, and N-NE sub-areas. Meta-gabbro does not appear to be present in the "central", E, and SE sub-areas, i.e. in the areas containing a large proportion of meta-psammitic rocks. The meta-gabbro bodies contain the same deformation features as the meta-sediments, and are thus interpreted to have intruded prior to the first phase of deformation (**D_1**). Hence, the intrusions probably occurred at the oceanic stage (see below). A syenite body of uncertain tectonostratigraphic status is present in the southern part of the study area.

Regional investigations

Reconnaissance work outside the study area, along the roads on the peninsula located between the fjords of Høydalsfjorden and Eikefjorden/Solheimsfjorden (**Fig. 2.4**), and on the peninsula situated between Høydalsfjorden and Førdefjorden/Brufjorden, have revealed mostly green- and grey-coloured sequences of interbedded meta-greywacke and meta-semipelite ("turbiditic layering"), and gradually changing meta-psammitic to meta-semipelitic and meta-pelitic rocks. Large occurrences of meta-psammitic rocks are present (i) on the islands of Alvora and Stavøya in Høydalsfjorden, where the meta-psammites show gradual primary transitions into meta-semipelitic rocks and meta-greywackes, and (ii) at Sandvika on the peninsula located between Høydalsfjorden and Eikefjorden/Solheimsfjorden. Conclusive evidence of ophiolite fragments like pillow lavas, sheeted dykes and ultramafics have not been observed. The only place where ophiolite fragments appear to occur, is at Stavang at the westernmost tip of the peninsula between Høydalsfjorden and Førdefjorden/ Brufjorden (**Fig. 2.4**), where Kildal (1970) reported that meta-ultramaphic bodies are present. These bodies are, however, located outside the study area, and have not been studied in the present thesis.

Interpretations of depositional environment

Interpretations in previous works: Based on extensive work in the Solund and Stavfjorden regions during the 1970s and 1980s, Furnes et al. (1990) formally established the Solund–Stavfjorden Ophiolite Complex (S–SOC), dated to **443 +/- 3 Ma** (Dunning & Pedersen 1988). From the same region they identified an oceanic-stage sedimentary cover sequence to the ophiolite, termed the Stavenes Group, which was found to consist of three units: the Heggøy Formation, the Hersvik Unit and the Smelvær Unit. Of these, the Heggøy Formation was seen to overlie the S–SOC with a primary depositional contact. On the Heggøy island itself, the formation was reported to consist of **c. 1000 m** of predominantly calcareous meta-greywacke, hosting numerous intrusive bodies and pillow lava horizons. Accordingly, the formation was interpreted to represent the oceanic-stage primary sedimentary cover to the ophiolite. The stratigraphic position of the other two units was reported uncertain, as no contacts to the ophiolite or cover rocks were exposed. The paper of Furnes et al. (1990) was — in addition to being based on the extensive investigations of the Solund and Staveneset regions — also founded on a few days of reconnaissance in the Høydalsfjorden region. On the bases of some lithological similarity, corresponding geochemistry, similar tectonostratigraphic position and a relatively moderate distance of geographical separation, Furnes et al. (1990) tentatively assigned the rocks of the present Høydalsfjorden Complex (HC) to the Heggøy Formation, and thereby made the HC part of the primary sedimentary cover to the Solund–Stavfjorden Ophiolite Complex.

Based on geochemistry and geological relationships, Furnes et al. (1990) concluded that the rocks in the present HC developed in a marginal basin between a continental margin on one side, and a volcanic island arc — situated above an active subduction system — on the other side. Furnes et al. (1990) tentatively suggested that a present-day analogue for the environment in which the rocks of the HC were formed, might be found in the area between Burma and Sumatra, i.e. in the N-S oriented Andaman Sea region of the Indonesian Arc system. The Andaman Sea is a marginal basin located between the N-S oriented continental margin of the Malay

Peninsula in the east, and a N-S directed volcanic island arc in the west, the Andaman and Nicobar islands. The basin is formed as a result of the oblique subduction of the Indian Plate beneath the Burma Plate, creating several ridge-transform segments in the marginal basin. The transform faults off-setting the ridge segments have orientations varying from NW-SE in the south via NNW-SSE to N-S in the north, i.e. the continental margin and a large portion of the transform faults are both oriented roughly N-S. Such parallelism between transform faults and a continental margin might turn the transform faults into obduction planes during later collision (Furnes et al. 1990). In the Sunnfjord region, the obduction of the outboard terranes might originally have been initiated along similar fault structures. The ridge segments have essentially SW-NE orientations (Furnes et. al. 1990).

In the original marginal basin where the rocks of the HC formed, the oceanic crust, developing during the active spreading in the basin, was gradually burried by a primary cover of sediments. Sediments were supplied from both the continent and the arc (and possibly from “oceanic islands” as well, see Furnes et al. 1990), and the turbiditic greywackes probably reached far out into the basin. The intrusive meta-gabbros may have originated from the magmatic source at the spreading ridge in the basin, and there intruded an unlithified primary cover of turbidite deposits and other "wet" sediments — i.e the intrusions occurred at the oceanic stage. As mentioned above, such basic intrusions into "wet" sediments have been described by e.g. Einsele (1982, 1985) from the present-day Bay of California. The large areas of meta-psammitic rocks in the "central", E, and SE sub-areas of the study area, which contain no meta-gabbro intrusions, could possibly have been situated at, or nearer to, the continental margin away from the magma source. The greenschists of the HC contain a high proportion of basic volcaniclastic material, indicating that this rock may have been formed near a basic volcanic source.

Discussion of alternative interpretations: In the following, two models will be suggested as alternatives to the above referred model where the HC-rocks have been correlated with the "cover sequence to the Solund–Stavfjorden ophiolite":

As a first alternative, the rocks in the HC may also be correlated with the ***Sunnfjord Melange*** which has been described from the Staveneset region (Osmundsen 1990; Andersen et al. 1990; Alsaker 1991; Alsaker & Furnes 1994; Osmundsen & Andersen 1994), located west of the Kvamshesten Devonian Massif. The melange, which is situated below the Solund–Stavfjorden Ophiolite Complex, was formed during the obduction of this complex, and is thus classified as an *obduction melange*. In the Staveneset region, the Sunnfjord melange displays features that are similar to the HC-rocks. For example, the Sunnfjord melange contains meta-greywacke and other meta-sediments (Andersen et al. 1990; Alsaker & Furnes 1994; Osmundsen & Andersen 1994) of the same type as in the HC. On the other hand, differences also exist. According to the descriptions of Andersen et al. (1990) and Alsaker & Furnes (1994), the Sunnfjord melange does not, for instance, appear to contain meta-gabbro bodies that have intrusive contacts towards surronding meta-greywacke. Furthermore, easily identifiable melange-type “mega-blocks” have not been observed in the HC, although they may still be present in the HC without being recognised because of the extensive Quaternary cover. It cannot, for example, be excluded that the “possible HC conglomerates” (Ch.4.2.1.2) are in fact “mega-blocks”. These similarities and differences may suggest that those parts of the HC-rocks (in the thesis area) that do not contain intrusive meta-gabbro bodies, might be part of an obduction melange similar to the Sunnfjord melange, instead of constituting the cover to an ophiolite. At Staveneset, Alsaker & Furnes (1994) reported that olistoliths (“mega-blocks") of basalt may have a length of up to **3 km**. If the Sunnfjord Melange are present in the thesis area, it is theoretically possible that those areas of the HC that contain meta-sediments intruded by gabbro, may themselves also be mega-olistoliths within a melange, i.e. making the whole HC a melange – although this hypothesis cannot be proved due to the low degree of exposure in the field area. Turbidites intruded by gabbro might have formed along the spreading ridge or near volcanic islands – on the condition that a turbidite-generating slope was sufficiently close. Also the Kvangagjelet Syenite may be considered in the light of the melange alternative. Although it is at present uncertain whether the Kvangagjelet syenite belongs to the HC or to the Håsteinen Devonian Massif, assigning it to the HC gives two possibilities: the syenite would be either (i) a mega-sized olistolith in a melange, or alternatively, (i) intrusive into the meta-sediments. This is further discussed later (Sect. 5.4.5).

As a second alternative, the HC rocks may be correlated with the ***Kalvåg Melange***, which is situated at the islands of Frøya and Bremanger (Bryhni & Lyse 1980; Bryhni et al. 1981; Bryhni & Lyse 1985; Ravnås 1991; Hartz 1992; Hartz et al. 1994; Steen 1994; Ravnås & Furnes 1995; Steen & Andresen 1997). The Kalvåg Melange has been interpreted as a bedded *olistostromal sedimentary melange* that was formed by submarin gravitational sliding and slumping of un- or partly-consolidated sediments down a tectonically controlled slope that persistently experienced mass-failure processes, and that was probably situated at the side of a volcanic arc that faced a marginal basin in which the S–SOC was formed (Ravnås & Furnes 1995). During the subsequent Scandian orogenic phase, the melange was little deformed. The Kalvåg Melange, which is a sedimentary melange *sensu stricto*, thus differs from the Sunnfjord Melange, which is a strongly deformed obduction melange. The Kalvåg Melange comprise features that can also be found in the HC, such as: **(1)** normally-graded, sandy Bouma-type T_{abc} and T_{ab} turbidites (see photo Fig. 2D in Ravnås & Furnes 1995); **(2)** beds being "chaotically" folded due to syn-sedimentary folding (see photo Fig. 9B and Fig. 9D of Steen & Andresen 1997), resembling the “disturbed “ rocks in the present thesis (Sect. 4.2.1.3); **(3)** breccia-resembling rocks stemming from: **3a)** a *small to moderate* degree of disintegration of an olistolith consisting of “interbedded sandstone and mudstone” (see photo Fig. 8A of Steen & Andresen 1997), and a *similar* degree of disintegration of an olistolith consisting of “shallow-marine deposits” (see photo Fig. 2F in Ravnås & Furnes 1995); **3b)** a *strong* degree of disintegration of an olistolith of bedded chert (see photo Fig. 9C of Steen & Andresen 1997); **3c)** a fully developed sedimentary breccia with no remaining bedding, thus forming a clast-rich debris-flow deposit being part of the melange "*groundmass*" (see photo Fig. 2G in Ravnås & Furnes 1995), which developed from less disintegrated beds (see photo Fig. 2F in Ravnås & Furnes 1995), resembling the «disrupted rocks» of the present thesis, (Sect. 4.2.1.3); **(4)** presence of conglomerates (see photo Fig. 2H in Ravnås & Furnes 1995).

At the present stage, the HC can thus be interpreted in three different ways: the rocks are either: **(1)** part of the primary sedimentary cover-sequence to an ophiolite which formed in a marginal basin/back-arc environment, and where the cover-sediments were intruded by gabbros originating in the subjacent oceanic crust; the ophiolite probably corresponding to the Solund–Stavfjorden Ophiolite Complex (Furnes et al. 1990); **(2)** part of an obduction melange, where the different rock-"types" may be mega-olistoliths in a manner possibly corresponding to the Sunnfjord Melange (Andersen et al. 1990); or **(3)** part of a sedimentary olistostromal melange, where the different rock-"types" may be mega-olistoliths in a manner similar to the Kalvåg Melange (Ravnås & Furnes 1995; Steen & Andresen 1997).

4.3 STRUCTURAL DEVELOPMENT OF THE HØYDALSFJORDEN COMPLEX (HC)

4.3.1 GENERAL

Section 4.3 contains a brief review of some main features of the structural development of the Høydalsfjorden Complex, including (i) very brief descriptions of the tectonic structures in the HC, and (ii) orientation analysis (stereographic projections) of the most important structural elements in the HC. The object of the chapter is to reveal which structures in the HC that *predate* the deposition of the Devonian sediments (i.e. structures cut by the sub-Devonian unconformity), and which structures that *postdate* the Devonian deposits. The question of whether the top-to-the-west movements that produced the myloniti-sation of the Eikefjord Group has also effected the HC rocks, is discussed. A comparison will be made between the **S_2**-mylonites of the Eikefjord Group and **S_3**-mylonites of the HC, as well as between the **F_3**-folds of the Eikefjord Group and **F_2**-folds of the HC.

As mentioned above, the Høydalsfjorden Complex in the study area has been divided into 8 sub-areas (**Plate 3**) to allow detection of potential local variations in the orientation of the structural parametres. The 8 sub-areas are "named" according to compass directions to the respective areas as if seen from central parts of the study area, starting in the west-northwest (WNW) and going clockwise around the Håsteinen Devonian Massif. The 8 areas are: (1) WNW, (2) N and NE, (3) E, (4) SE, (5) S and SW, (6) WSW, (7) W, and (8) "Central" (the Osen meta-psammite inlier). The same sub-areas were also used during the rock descriptions (Sect. 4.2) to indicate the positions of geographical names on the main map (**Plate 1**).

It is reminded that the numbering of deformation phases in the present thesis relate exclusively to each tectonostratigraphic unit and only reflects the structural history within this particular unit. Based on the present structural interpretation, this means that D_1-structures in one unit (e.g. the Høydalsfjorden Complex) do not a priori correspond to D_1-structures in other units (i.e the Eikefjord Group or the Håsteinen Devonian Massif), and so on with D_2 , etc.

4.3.2 ABSENCE OF 3-D CONTROL IN THE HØYDALSFJORDEN COMPLEX

In the Høydalsfjorden Complex, no marker lithologies or stratigraphy exist that can reveal the three dimensional (3-D) rock architecture. Instead the rocks show only very gradual transitions between regions that are more meta-psammitic, more meta-pelitic or more bimodal (meta-greywacke interbedded with meta-semipelite) (see Sect. 4.2.1). As will be described below, the HC rocks in the study area, as well as in neighbouring areas around Høydalsfjorden, have been folded into WNW-ESE to NW-SE trending **F_2**-folds that are commonly seen on outcrop-scale (meter-scale). The lack of marker horizons or a stratigraphy, however, implies that the large-scale fold-shapes (i.e. **100 meter-scale** or **km-scale**) are not delineated in the field. This means, for example, that the lithological layering seen to be folded at one outcrop, cannot be correlated to layering at another outcrop some distance away, which also implies that it is impossible to know whether one is moving upwards or downwards in the stratigraphy as one traverses the folded terrane in e.g. the N-S direction. In addition, the vergence of mesoscopic folds is seen to vary from outcrop to outcrop. Since it is therefore, all in all, impossible to relate outcrop folds to possible larger-scale folds, it is not possible to achieve control of *overall vergence* of higher order folds; and any possible large-scale structures cannot be unravelled. Therefore, in the profiles (**Plate 2**), the HC is drawn without ”fold-structures”, and for the same reasons the folds in the **Road Logs 1–6** cannot be used to aquire information on *overall vergence* of the folds (see Sect. 4.3.4.2).

4.3.3 POSSIBLE D_0-STRUCTURES

The primary lithological layering in the meta-sediments, e.g. the interlayering of turbiditic meta-greywacke and meta-semipelite, is termed **S_0**. Possible syn-sedimentary or pre-lithification slumping and sliding is classified as **D_0**-deformation.

The "disturbed" (not "disrupted"!) layering that was described in Sect. 4.2.1.3 may possibly be interpreted as a result of syn-sedimentary slumping or sliding processes (**Fig. 4.13**). The structures are partly chaotic with, for example, layers of meta-greywacke terminating towards each other at variable angles. The odd geometries indicate that the structures have probably not been produced by ordinary tectonic folding-processes in lithified rock. Lack of descrete, brittle fault zones in the rocks also exclude that fault displacements in lithified rock was responsible for the structures. Although soft-sedimentary deformation seems to have operated, it is difficult to find conclusive evidence for this type of deformation. This is due to lack of exposures; later overprinting tectonic deformation of the rocks; and particularly the presence of the **D_2**-"disruption" features described in Sect. 4.2.1.3.

4.3.4 D_1-STRUCTURES

4.3.4.1 S_1-FOLIATION

S_1-Cleavage

The **S_1**-cleavage is the dominant tectonic fabric in the HC, and is effecting all rock types in the complex (except where zones of **D_3**-mylonites are present, see below). The **S_1**-fabric always appears to be parallel to bedding (**S_0**) where bedding is present (**Fig. 4.3**, **4.6**) **—** with exception of a few instances where small deviations have been observed. Microscopically, the cleavage in the meta-sediments is defined by parallel orientation of longest dimension of grains of white mica, chlorite, quartz, feldspar, epidote, etc. The **S_1**-foliation is normally a result of mainly coaxial deformation, but shear fabrics may occur. The orientation of the **S_1**-fabric is controlled by the later **F_2**-folds, and the **S_1**-fabric defines the limbs of these folds (see Sect. 4.3.5.2).

Discussion/interpretation: The parallelism between the **S_1**-foliation and the **S_0**-bedding in the HC may theoretically have been accomplished in four different ways:
(1) through isoclinal **F_1**-folding of **S_0**;
(2) through predominantly bedding-parallel non-coaxial deformation
(3) through predominantly coaxial contraction at right angle to the bedding; and
(4) trough combinations of (2) and (3).

Alternative **(1)** (isoclinal folding) is problematic, since conclusive examples of isoclinal **F_1**-folds have not been observed. However, since the exposures are limited, the alternative cannot be excluded. Alternative **(2)** (bedding-parallel non-coaxial deformation) is possible since **D_1**-shear fabrics are present. At outcrop-scale, it is difficult to find conclusive evidence that such penetrative shear movements have been responsible for the complete **S_1**-fabric, but shear may have played a role locally. Alternative **(3)** (bedding-orthogonal coaxial contraction) is also possible, since the **S_0**-/**S_1**-parallelism also exists in areas which apparently have not been subjected to any obvious shear. Compaction may have contributed to this type of deformation. Alternative **(4)** (combinations of (2) coaxial and (3) non-coaxial deformation) is possible, since the combinations is more common in orogenic deformation than the "end-members".

Although isoclinal folds have not been observed, all four factors may thus have contributed to the **S_0**-/**S_1**-parallelism within the study area. All factors would comply with the model of thrust obduction of the outboard terrane onto the Balto-Scandian platform (Sect. 4.3.5), since any orogenic process generally involves components of coaxial and non-coaxial deformation in variable amounts in space and time.

As will be discussed later (Sect. 4.3.6), the HC obviously has a Scandian contractional (eastward) thrust history related to the emplacement of the complex onto the Balto-Scandian craton, and the **D_1** structures may be related to this deformational phase. As will also be treated later (Sect. 4.4), the HC experienced a peak metamorphism corresponding to the upper part of the *lower* greenschist-facies. Regarding the extensional mylonites in the Eikefjord Group, the latter were described (Sect. 3.5.2) as having retrograded to *middle* or *lower* greenschist-facies mylonites at the last stages of their shear movements. When comparing the two metamorphic developments, this implies that the late-stage *extensional* metamorphic conditions in the Eikefjord Group were fairly similar to the main *contractional* metamorphism in the HC, although the two events occurred at different points in time.

4.3.4.2 POSSIBLE F_1-FOLD

F_1-folds, i.e. folds having the **S_1**-foliation as axial planar cleavage, have generally not been observed in the study area. The only exception from this is one single fold observed at a locality situated about **100 m** to the south of the point where the county road 542 (fylkesveg 542) crosses the river Pollevikelva (sub-area: WNW). At this locality, the main **S_1**-cleavage appears to have an axial planar position in a folded "turbiditic" meta-greywacke layer that is surrounded by layers of meta-semipelite. However, the exposure is limited, neighbouring layers do not appear to have been folded in a similar way, and tectonic shear does not appear to have been dominating at the locality. Elsewhere in the same area, "chaotic" disturbances of the layering have been interpreted as a result of syn-sedimentary slumping (Sect. 4.2.3). It is therefore also possible that the present fold is an **F_0**-fold and not an **F_1**-fold, and that the **S_1**-cleavage received an orientation which accidentally gives the impression that the cleavage is axial planar to the fold.

As noted above, the complete absence of marker lithologies makes it impossible to tell whether the composite **S_0**-/**S_1**- fabric described above (Sect. 4.3.3.1) was a result of large-scale isoclinal **F_1**-folding.

4.3.5 D_2-STRUCTURES

4.3.5.1 S_2-CLEAVAGE

The **S_2**-cleavage is generally not penetrative or continuous. Where present, it is defined by discrete kink bands, or in certain outcrops by a spaced crenulation cleavage, that crosses the **S_1**-cleavage. The spacing of the crenulation cleavage is typically **1–2 cm** (**Fig. 4.33**). The space between the kink bands may be up to **10 cm** (**Fig. 4.34**). The kink bands may display conjugate sets, especially in meta-psammitic lithologies. Along strike, the kink bands commonly form an anastomosing pattern (**Fig. 4.35**). The **S_2**-cleavage generally has a W-E to WNW-ESE strike, and steep dips both towards the N–NNE and S–SSW. Locally, where the space between the kink bands is narrow, they approach the appearance of the spaced crenulation cleavage, indicating that the two types of structures are related. Microscopically, the **S_2**-fabric does not display growth of new minerals; it merely reorients the minerals defining the **S_1**-planar fabric. Domainal removal of quartz through pressure solution may be present in the kink bands.

4.3.5.2 F_2-FOLDS

The **F_2**-folds are the dominant fold structures in the HC and fold the composite **S_0-/S_1**-fabric. The folds may be of both chevron type and more "curved" type. The chevron folds appear to occur in the more quartzo-feldspatic rocks (**Fig. 4.36**), and the curved ones in the rocks consisting of interchanging meta-greywacke/meta-semipelite (**Fig. 4.37**), as well as in the undifferentiated meta-semipelites. **D_2**-structural elements are shown in stereographic plots corresponding to the sub-areas in the WNW **(Fig. 4.38**), N-NE (**Fig. 4.39**), E (**Fig. 4.40**), SE (**Fig. 4.41**), S-SW (**Fig. 4.42**), WSW (**Fig. 4.43**) W (**Fig. 4.44**), and the "central" Osen meta-psammite inlier (**Fig. 4.45**). The **F_2**-folds are clearly cut by the sub-Devonian unconformity (see Sect. 5.3). The formation of the «disrupted rocks» described in Sect. 4.2.1.3, was probably, to a considerable extent, produced during the **F_2**-folding, as a result of a difference in competence between the interlayered beds of "turbiditic" meta-greywacke and meta-semipelite. The **F_2**-related “disruption” has also affected more homogeneous meta-psammites (e.g. at Osknoltren in the N-NE sub-area), which have less internal competence contrasts, indicating that the “disruption” also affected rocks with less different rheologies.

Orientations of F_2-fold axes

The **F_2**-fold axes in the different sub-areas have been plotted separately to allow local variations to be detected. However, in all 8 sub-areas, the FAs are oriented with WNW-ESE directed azimuths, and in all areas except at Gravaneset (subarea: W), the plunges generally are in the range of **0–50°** in both directions (**Fig. 4.38a**, **4.39a**, **4.40a**, **4.41a**, **4.42a**, **4.43a** and **4.45a**), i.e. with an average plunge of **c. 25°**.

A separate plot is presented for the small Gravaneset area (**Fig. 4.44a**). As usual, the FAs are distributed along a WNW-ESE vertical great circle, but the variation in plunge is larger than normal for such a *small area*, indicating that the folds might be highly noncylindrical. In the forthcoming Sect. 5.5.3.3, on the structural development of the Håsteinen Devonian Massif, these FAs of the Gravanest HC rocks will be compared to the parasitic folds in the Gravaneset Devonian rocks just above the unconformity.

Orientations of axial planes of F_2-folds

The axial planes are essentially oriented with WNW-ESE strike, and steep to moderate dips in both directions, causing the axial planar poles to be distributed in the NNE and SSW parts of the plots (**Fig.**

4.38a, **4.39a**, **4.40a**, **4.41a**, **4.43a**, **4.44a** and **4.45a**). In the subareas in the WNW, N-NE and W, the axial planes are mainly dipping SSW-wards, with a majority of the poles located in the NNE area of the stereo plots (**Fig. 4.38a**, **4.39a**, and **4.44a**). In other plots, the axial planes dip both to the SSW and NNE, in more equal numbers.

Orientations of the limbs of F_2-folds

The orientation of the composite S_0-/S_1-fabric is controlled by the F_2-folding (**Fig. 4.38b**, **4.39b**, **4.40b**, **4.41b**, **4.42b**, **4.43b**, **4.44b** and **4.45b**). The limbs to the F_2-folds are generally oriented with a WNW-ESE directed strike, and steep to moderate dips towards both the NNE and SSW. In the sub-areas in the WNW, N-NE, E, W, and in the "central" Osen meta-psammite inlier, the majority of the limbs recorded dips southwards (**Fig. 4.38b**, **4.39b**, **4.40b**, **4.44b**, **4.45b**).

Fold axes, axial planes and limbs with deviating orientations. (Possible F_3-folds)

In four of the sub-areas, (WNW, N-NE, E and WSW), a few FAs have a trend directed NE-SW, i.e. roughly orthogonal to the common F_2-trend. Likewise, a few axial plane poles and fold limb poles are located in the NW or SE parts of the stereoplots, which is also orthogonal to the common orientation, but consistent with the deviating FAs. These deviating structural elements are very rare in the study area, and are possibly a result of a phase of F_3-folding (see Sect. 4.3.7.1).

The Road Logs

The style of the F_2-folds are shown in the **Road Logs** in Appendix B, and the structural measurements of the various elements are placed on the logs. **Road Logs no. 1** and **2** transect the WNW sub-area. It can be seen in the logs that the folds have no dominant vergence. (**Fig. 4.37** is taken from stations no.11–12 in **Road Log no. 2**). **Road Log no. 3** transects the WSW sub-area. Note the meta-gabbro bodies in the section. **Road Log no. 5** goes through the E sub-area, and **Road Log no. 6** crosses the SE sub-area. On all these sections it can be seen that the F_2-fold vergence varies from place to place, according to variations in the orientations of the axial planes and the S_0/S_1-limbs. This, and the lack of marker horizons, makes it impossible to deduce the large-scale fold structure.

The profiles

In the four constructed profiles transecting the study area (**Plate 2**), the HC is drawn to show no fold patterns or other structural geometries. This is again due to the complete lack of structural markers in the field, which makes it impossible to unravel the large-scale three dimensional architecture of the rocks. Likewise, vergence of mesoscopic folds cannot be related to larger-scale structures, and therefore vergence-indications of the F_2-folds have not been shown in the profiles.

4.3.5.3 COMPARISON BETWEEN HC F_2-FOLDS AND EG F_3-FOLDS

Orientation and relative age of HC F_2-folds and EG F_3-folds

As seen from the stereo plots (e.g. compare **Fig. 4.39a** and **Fig. 3.29**), the **F_2**-fold axes in the Høydalsfjorden Complex (HC) and the **F_3**-fold axes in the Eikefjord Group (EG) (Sect. 3.4.3) are oriented with essentially the same W-E to WNW-ESE trends. The amount of plunge is also essentially overlapping, although the EG **F_3**-folds tend to have somewhat shallower plunges, and the HC **F_2**-folds a larger spread. Furthermore, in both groups the majority of the axial planes tend to strike W-E to WNW-ESE, although with highly variable dips (e.g. compare APs to HC **F_2**-folds, **Fig. 4.40a** and APs to EG **F_3**-folds, **Fig. 3.29a**).

Although roughly similar in orientation, the **F_2**-folds in the HC cannot be correlated with the **F_3**-folds in the Eikefjord Group/NSDZ. The HC **F_2**-folds definitively pre-date deposition of the Devonian sediments, since the folds are truncated by the sub-Devonian unconformity **—** and the HC **F_2**-folds thereby also predate the folding of the Devonian deposits. Regarding the EG **F_3**-folds, these are folding extensional mylonites of the Nordfjord–Sogn Detachment Zone. The HC **F_2**-folds were formed during the Scandian top-to-the-*east* orogenic contraction (Sect. 4.3.6), whereas the EG **F_3**-folds were formed during the subsequent post-Scandian top-to-the-*west*, crustal-scale extension (Sect. 3.4.3).

The time relationship between the HC **F_2**-folds and the EG **F_3**-folds are illustrated in **Table 4.1**. (In addition, the table also shows the time relationships between all deformation phases within all three tectonostratigraphic units of the study area. References will be made to **Table 4.1** also in later sections and chapters, when time relationships between deformation phases are discussed).

Time limits for the formation of HC F_2-folds and EG F_3-folds

Scandian top-to-the-east nappe transport, which included the HC rocks, started at **~ 420 Ma** and continued until at least **408 Ma**, as documented by $^{40}Ar/^{39}Ar$-dating of muscovite from the décollement zone below the Jotun Nappe (Fossen & Dunlap 1989). This age of **408 Ma** may represent a theoretical youngest minimum age for the HC **F_2**-folds **—** i.e.: the HC **F_2**-folds formed prior to, or around **408 Ma** (**Table 4.1**).

If the HC **F_2**-folds were formed as late as **~ 408 Ma**, i.e. while the Scandian top-to-the-*east* transport still occurred, the formation of the folds would overlap in time with the **~ 412–405 Ma** age ($^{40}Ar/^{39}Ar$ on amphibole) of the lower amphibolite facies top-to-the-*west* mylonites of the Mode-II/NSDZ (see Sect. 3.5.2). Johnston et al. (2007b) suggested that the NSDZ started to move at **410 Ma**, which was based on the following reasoning: Sm-Nd dating of garnet rims gave this age of **~410 Ma**, which was interpreted to mark the end of pre-NSDZ peak upper amphibolite facies metamorphism. After this, the lower amphibolite–greenschist facies NSDZ shear could start. If the NSDZ was formed at **410 Ma**, this would partly overlap with formation of Scandian $\mathbf{F_2}$-folds at **408 Ma**.

The $\mathbf{F_3}$-folding of the Eikefjord Group $\mathbf{S_2}$-mylonites (of Mode-II/NSDZ), took place some time after the formation of the mylonites had started. The timing of the formation of the mylonites is a bit unclear due to an apparent mismatch between $^{40}Ar/^{39}Ar$ ages and the interpretation that the EG/NSDZ is a Mode-II zone. (See discussion in Sect. 3.5.2). However, based on fieldwork, it is established in the literature (Fossen 1992, 2000; Milnes et al 1997) that the top-to-the-west Mode-I movements of the orogenic wedge, on the décollement zone below the Jotun Nappe, occurred prior to the Mode-II movements on the NSDZ. The Mode-I movements were dated by $^{40}Ar/^{39}Ar$ on *muscovite*, to **402–394 Ma** (Fossen & Dunlap 1998). The subsequent Mode-II movements were probably initiated slightly prior to the **394 Ma** age, i.e. slightly prior to the last stages of the Mode-I decollement movement (**Table 4.1**). Accordingly, the $\mathbf{F_3}$-folding of the $\mathbf{S_2}$-mylonites of the Eikefjord Group/NSDZ probably took place later than **394 Ma**.

Final remark

In summary, the HC $\mathbf{F_2}$-folds formed in the time interval **420–408 Ma**, during Scandian top-to-the-east movements, whereas the EG $\mathbf{F_3}$-folds probably formed subsequent to **394 Ma**, i.e after initiation of the Mode-II/NSDZ/EG movements that led to development of the EG $\mathbf{S_2}$-mylonites (**Table 4.1**). The HC $\mathbf{F_2}$-folds are cut by the unconformity below the Håsteinen Devonian Massif. The initiation of basin formation and deposition of the Håsteinen sediments was probably associated with the formation of the Mode-II/NSDZ that produced the EG mylonites. If the deposition of Devonian sediments started somewhat later than the initiation of the detachment zone, it is theoretically possible that the EG $\mathbf{F_3}$-folds started to form shortly before deposition of the Håsteinen sediments. However, it appears more likely that the EG $\mathbf{F_3}$-folds were formed during the deposition of the Håsteinen basin, i.e. at some stage during the Mode-II movements on the NSDZ. The $\mathbf{F_3}$-folding of the EG mylonites probably took place contemporaneously with the $\mathbf{F_1}$-folding of the Håsteinen Devonian Massif (**Table 4.1**). In conclusion, the Eikefjord Group $\mathbf{F_3}$-folds formed much later than the $\mathbf{F_2}$-folds of the Høydalsfjorden Complex.

4.3.6 INTERPRETATION OF D_1- AND D_2-STRUCTURES AS SCANDIAN

The absolute *maximum* age for the deformation process that yielded the HC $\mathbf{D_1}$- and $\mathbf{D_2}$-structures is defined by the age of formation of the rocks. This age is found by correlation of the HC rocks with similar rocks at Staveneset, i.e. with the primary sedimentary cover to the Solund–Stavfjorden Ophiolite Complex. Such a correlation is suggeted by Furnes et al. (1990). Diorite in this ophiolite complex has been radiometrically dated to **443 +/- 3 Ma** by the U/Pb method (Duning & Pedersen 1988), and this age is interpreted to be the age of formation of both the ophiolite and the sedimentary cover (Furnes 1990), i.e. representing the absolute maximum age for the deformation. However, the rocks of the HC may also be correlated with the Sunnfjord Melange (a view also held by Harald Furnes, pers. com.), and this provides a possibility to further refine the maximum age for the deformation: The Sunnfjord Melange was formed during the obduction of the Solund–Stavfjorden Ophiolite Complex onto the continental margin (Andersen et al. 1990). As reported by Andersen et al. (1990), the rocks underlying this melange contain fossils of probably **Wenlock age** (**428–423 Ma**, time-scale of Gradstein & Ogg 1996; and Gradstein et al. 2004), and this implies obduction of the melange subsequent to the fossil formation, i.e. obduction during or after **Wenlock**. This age of obduction represents a *younger* maximum age for the formation and deformation of the melange and the overriding Solund–Stavfjorden Ophiolite Complex – i.e. for the **D$_1$**- and **D$_2$**-deformation of the HC. In summary, this means that the obduction and related deformation of the rocks in the Høydalsfjorden Complex occurred during or later than **Wenlock** (**428–423 Ma**).

An absolute *minimum* age (i.e. the youngest minimum age) for the deformation of the Høydalsfjorden Complex is represented by the unconformity below the Middle Devonian <u>Håsteinen</u> Massif which cuts the **F$_2$**-folds in the HC; and also by the unconformity below the Middle Devonian <u>Kvamshesten</u> Massif which unconformably truncates the Sunnfjord Melange at Staveneset (Osmundsen et al. 1998). However, by considering eclogite formation in the Western Gneiss Region, also this minimum age may be substantially better constrained: The eclogites in the Western Gneiss Region formed in the interval **420–400 Ma**, corresponding to **Late Silurian–Early Devonian times** (Griffin & Brueckner 1980; Griffin et al. 1985; Carswell et al 2003a). These high-pressure rocks reflect westward subduction of the Western Gneiss Region (i.e. the Baltic margin) below the Laurentian craton. This subduction implied that the orogen had reached the stage of full continent-continent collision, which means that all marginal basins and other outboard terranes (including the Høydalsfjorden Complex rocks) outside the continental margins would have been, or would be in the process of being obducted onto this margin. By the time of the formation of the first eclogites, sometimes **around 420 Ma**, the obduction of the outboard terranes had probably already started. This implies that the obduction probably took place during late Silurian times, i.e. in the **Ludlow–Pridoli** (**423–417 Ma** according to the time-scale of Gradstein & Ogg 1996; **423–416 Ma**, Gradstein et al. 2004) and this orogenic phase might have produced the **D$_1$**- and **D$_2$**-deformation in the HC. The best estimate of the *oldest minimum* age of the HC deformation is thus **Ludlow–Pridoli**, and Andersen et al. (1990) indicated a roughly similar time frame, by suggesting that the obduction occurred during **Wenlock** times. Nevertheless, as briefly mentioned above (Sect. 4.3.5.3), the **D$_1$**- and **D$_2$**-deformation might theoretically have originated at any time during the Scandian top-to-the-east nappe transport. From the decollement zone below the Jotun Nappe, Fossen & Dunlap (1998) showed that this eastward transport lasted until at least **408 Ma** (see **Table 4.1**). Furthermore, the low metamorphic grade of the HC rocks, i.e. lower greenschist-facies, shows that the rocks were always positioned at a shallow level in the nappe stack, probably experiencing a maximum temperature of **350–400 °C**, which with a thermal gradient of **30 °C/km** would correspond to a depth of **~12 km**. It is reasonable to assume that the most intense deformation of the HC occurred during obduction of the rocks, i.e. implying that the **D$_1$**- and **D$_2$**-deformation was probably associated with this obductional phase.

In conclusion, these time frames imply that the **D_1**- and **D_2**-structures were presumably formed in the Late Silurian, possibly **Wenlock–Pridoli times**, as a result of the obduction of the outboard terranes onto the Balto-Scandian platform. However, it cannot be excluded that the deformation took place at later stages, since, as documented by Fossen & Dunlap (1998), Scandian top-to-the-east nappe transport lasted until at least **408 Ma** (i.e. until at least **Praghian** of the Early Devonian, time-scale of Gradstein & Ogg 1996; and Gradstein et a. 2004). The above events occurred during the major Scandian phase of the Caledonian orogeny (e.g. Andersen et al. 1990; Fossen 1992), a phase occurring in the time interval **Late Silurian to Early Devonian**. This phase represented the major continent-continent collision between the Laurentian and Balto-Scandian cratons at **425-400 Ma**.

4.3.7 D_3- STRUCTURES

4.3.7.1 S_3-MYLONITES (TOP-TO-THE-WEST)

The **S_3**-mylonites of the HC form a very characteristic rock type that is normally easy to recognise in the field. The mylonites may be observed several places in the study area, for example **(i)** between the Devonian rocks of Vikafjell and the Sunnar Fault (sub-area: N-NE), **(ii)** east of Nonsnova (sub-area: E), and **(iii)** in the road-cut east of Svardal (sub-area: SE). In the following presentation, area **(i)** will be described in some detail, from both Sunnarviken and higher up on the mountainside.

(i) In the north-dipping mountainside located between Vikafjell and the Sunnar Fault (sub-area: N-NE), zones of mylonite appear to form morphological “steps” in the steep terrain, “steps” that are defined by vertical cliffs of mylonite being up to some tens of meters high, and with top-surfaces that are more flat-lying. The top surface of each “step” defines a shelf that may be traced laterally for **several kms** along the mountainside, the shelf being gently inclined towards WNW. Shear sense in the mylonites (see below) appear to be roughly *top-to-the westnorthwest,* for simplicity written *top-to-the-west* in the following discussion. The mylonites may conveniently be observed in the southern part of Sunnarviken (sub-area: N-NE), in the large road-cut going westwards from the onshore segment of the Sunnar Fault (**Plate 1**). Generally, the mylonites are recognised by their characteristic fine-grained appearance, usually monotonous dark to medium, grey to greenish-grey colour and very strong fabric. Occasionally, a moderate lithological banding or "lamination" is present, and on rare occasions, this banding may be quite pronounced for this mylonite, as observed for example at an outcrop northeast of the summit of Vikanipa, near the small E-W oriented Devonian spur (sub-area: N-NE, at **ca 230 m.a.s.l.**) (**Fig. 4.32**). Shear-sense indicators are usually difficult to detect at the outcrops. This is due to a very homogeneous nature of the mylonites, with generally no porphyroclasts; and with little or no lithologic banding, implying a lack of markers to display shear bands, foliation boudinage, etc.

On the microscopic scale there is extensive growth of white mica and chlorite in the matrix, whilst feldspars display brittle behaviour. When feldspars are abundant, they may give the rock a “semi-cataclastic” appearance. Quartz grains may have a “porphyroclastic” appearance in the matrix (**Fig. 4.46a, 4.46b**). These grains have subgrain tails, and are also to a variable extent recrystallised into sub-grains, both along the rims and throughout. Quartz grains with “porphyroclastic” properties may also show shear-related disintegration into smaller pieces. At times, such disintegration have occurred along planar grain surfaces that may be slightly inclined to the surrounding average mylonite banding, and grain pieces in contact with each other are then displaced with a “normal sense” of movement (**Fig. 4.46a**). These planar surfaces indicate that such failure of quartz grains was initiated by brittle cracking. Nevertheless, the surfaces are almost always decorated by tiny quartz subgrains (**Fig. 4.46b**), yet preserving the planar shape of the surfaces (**Fig. 4.46a**). The subgrains indicate that crystal plastic processes operated simultaneously with the cracking, suggesting that the temperature during deformation may have been around **300 °C** (Passchier & Throuw 1996). The mylonite matrix, as seen in thin-section, commonly show a strong black-coloured banding resulting from “down-milling” to cryptocrystalline grain sizes that prevent light transmittance (**Fig. 4.46**).

East-northeast of the summit of Vikanipa (sub-area: N-NE), high up on the N-dipping mountainside (at **ca 375 m.a.s.l.**), a belt of mylonite is present, having a S-dipping and WNW-ESE striking foliation, and defining a zone **several tens of metres** thick and with a lateral E-W continuity of at least **200 m** in the mountainside. The rocks crop out in a steep cliff above an accessable shelf in the mountainside, and are exposed for most of the lateral **200 m**. Although shear sense was not obtained at this locality, the appearance of the rock seems to be identical to the one in Sunnarviken, both at the meso- and microscopic scale, and the rock is thus interpreted as a top-to-the-west $\mathbf{S_3}$-mylonite. The rock exposures can be traced westwards towards the Devonian/substrate contact, but the area defining **the last few metres** before the contact, as well as the contact itself, are covered by vegetation. Although the precise contact is not exposed, these mylonites must be cut by the sub-Devonian angular unconformity, since such an unconformity is present at the Håsteinen/substrate contact located **100–200 m** (map distance) south of the mylonite belt, in the area where the major mountain ridge crest goes eastwards from the major Vikanipa Devonian spur.

(ii) Also in the “sub-area: E”, at a locality to the south-southeast of the mountain Nonsnova, where the sub-Devonian primary angular unconformity is exposed, it can be clearly seen that the sub-Devonian unconformity cuts typical greyish HC mylonites of the same type as those in the mountainside between Vikafjell and Sunnarviken.

(iii) The exposures of mylonites in the road-cut east of Svardal (sub-area: SE) has a similar appearance to the other mylonites mentioned above. The Svardal rock is thus interpreted as an $\mathbf{S_3}$-mylonite, although it has not been studied in detail.

The above examples show that the HC contains top-to-the-west mylonites that are definitely **pre-Devonian** of age, i.e. pre-dating deposition of the Devonian sediments.

Kinematic indicators: The $\mathbf{S_3}$-mylonites in the road-cut that goes westwards from the onshore segment of the Sunnar Fault in Sunnarviken (**Plate 1**), are situated in the tectonostratigraphic lowermost parts of the HC. This mylonite outcrop has been investigated for sense of shear, to reveal whether the mylonites show top-to-the-west movements, which would be consistent with the movements in the mylonites of the neighbouring Eikefjord Group (Sect. 3.4.2), i.e. the Nordfjord–Sogn Detachment Zone. At the outcrop, a general lack of both lithologic banding and porphyroclasts in the monotonous dark-grey, fine-grained mylonites (see above), implied that good shear sense indicators could not be observed. From a **30 x 7 x 7 cm** oriented sample of a typical mylonite, taken in this road-cut at a point about **50 m** westwards from the contact to the Eikefjord Group, six thin-sections were investigated for shear-sense indicators. These sections clearly showed a ***top-to-the-west*** sense of shear (**Fig. 4.46**), i.e. similar to the shear direction of the Eikefjord Group. Indications of contractional top-to-the-*east* movements, related to the Scandian emplacement of the HC rocks onto the Baltic craton, were not observed in the mylonites.

In the mountainside between Vikafjell and Sunnarviken, mylonites are present near the small E-W oriented Devonian spur (at **ca 230 m.a.s.l.**), located northeast of the summit of Vikanipa. Again the sense of shear in the mylonite is difficult to deduce in the field, even though the rock is exceptionally well banded for this mylonite (**Fig. 4.32**). Nevertheless, the outcrop shows vague kinematic indicators, such as trains of tiny sigmoidal-shaped “lenses ” of whitish band material, up to **10 mm** long and **1–2 mm** thick, that appear to suggest top-to-the-west movement (lowermost part of **Fig. 4.32**), i.e. consistent with the result at Sunnarviken.

4.3.7.2. COMPARISON OF HC S_3-MYLONITES WITH EG S_2-MYLONITES

On the assumption that this top-to-the-*west* sense of shear is representative for the mylonitic HC rocks in the study area, it is of interest to consider whether these top-to-the-west HC **S_3**-mylonites are genetically related to the top-to-the-west **S_2**-mylonites in the Eikefjord Group. A simultaneous development of the two mylonites would have to imply that movements on the NSDZ, which produced EG **S_2**-mylonites, occurred *prior* to deposition of the Devonian sediments, since the HC **S_3**-mylonites are cut by the Devonian unconformity. The comparison will be based on some time limits for the shear movements:

Time limits for the formation of the Eikefjord Group (EG) **S_2**-mylonites (Mode-II of Fossen & Dunlap 1998) may be outlined as follows: The top-to-the-west movement of the entire orogenic wedge (Mode-I extension of Fossen & Dunlap1998), which occurred on the décollement zone, represents the first stage of top-to-the-west movement in the Caledonides of southern Norway, and occurred in the interval **402–394 Ma** (Fossen & Dunlap1998) (**Table 4.1**). During this movement, however, the underlying subducted eclogite-containing Western Gneiss Region experienced exhumation, leading to uplift also of the above-lying décollement zone, as well as the orogenic nappes. Eventually, at around **394 Ma**, the Mode-I décollement zone had lost its westward dip and become close to horizontal. Then the movements along the Mode-I zone came to a halt, and was replaced by the Mode-II extension, represented by the Nordfjord–Sogn Detachment Zone (NSDZ) (Fossen & Dunlap1998). The NSDZ cut through the nappe stack and formed the EG **S_2**-mylonites. Mode-II movements may have overlapped in time with the Mode-I phase, and the **402–394 Ma** Mode-I movements thus represents the earliest possible time of movements on the EG **S_2**-mylonites. It is, however, likely that the EG **S_2**/NSDZ/Mode-II movement was initiated at the last stages of, or shortly after, the time interval for the Mode-I extension (**402–394 Ma)** (**Table 4.1**).

Time frames for the development of the Høydalsfjorden Complex (HC) **S_3**-mylonites may be presented as follows: also for the HC **S_3**-mylonites, the earliest possible time of mylonite-formation would probably correspond to the **402–394 Ma** interval of the Mode-I extension. This implies that the HC **S_3**-mylonites could have developed also partly simultaneously with the Mode-II EG **S_2**-mylonites, which probably started to form at the last stage of, or just after the Mode-I phase. The HC **S_3**-mylonites reflect movements that took place at around **300 °C**. With a thermal gradient of **30 °C/km**, this implies that **~10 km** of overburden had to be removed by tectonic and/or erosional processes before the deposition of the Devonian sediments could occur, since the mylonites are truncated by the sub-Devonian unconformity. Although the HDM sediments are assumed to be of Middle Devonain age, (**391–370 Ma**, time-scale of Gradstein & Ogg 1996; **398–385 Ma**, Gradstein et al. 2004), the deposition did not necessarily start at the very beginning of the Middle Devonian, i.e. at **391/398 Ma**. This means that there would be some time available for the (i) formation of the HC **S_3**-mylonite at **10 km** depth, at around **402–394 Ma**, and for (ii) the subsequent erosion that with a rate of **1 mm /year** would bring the mylonites to daylight exposure within **10 million years**, and with a rate of **2 mm/year,** to daylight exposure within **5 million years**. The latter rate of erosion is possibly higher than realistic, and lack of time would mean that tectonic prosesses would have to contribute to the processes bringing ther HC **S_3**-mylonites to the surface where deposition of Devonian sediments took place. However, after the beginning of deposition of Devonian sediments that unconformably covered the HC **S_3**-mylonites, the extensional movements along the Eikefjord Group **S_2**-mylonites *continued* for a long time throughout the Middle Devonian period, now controlling the formation of, and deposition in the Devonian basin.

As will be discussed later (Sect. 4.4), the HC **S_3**-mylonites formed at a metamorphic grade lying just below greenschist-facies. The extensional **S_2**-mylonites in the Eikefjord Group were, however, in a previous section (Sect. 3.5.2) described as having been retrograded to *middle* or *lower* greenschist-facies mylonites at the last stages of their shear movements. When comparing the metamorphic development of the two, this implies that the metamorphic conditions during late-stage extensional movements in the Eikefjord Group were of higher temperature than the metamorphic conditions during extensional movements in the HC.

From the above discussions it appears that movements along the HC **S_3**-mylonites could only have occurred simultaneously with the *early-phase* movements on the EG **S_2**-mylonites (since the HC **S_3**-mylonites are cut by the Devonian unconformity), whereas the EG **S_2**-mylonites had a continued, long-lasting history associated with the formation of the Devonian basin (**Table 4.1**).

4.3.7.3. THE POTENTIAL FOR TOP-TO-THE-EAST STRUCTURES NEAR THE S_3-MYLONITES

The HC obviously has a Scandian contractional (eastward) history related to the thrust emplacement of the complex onto the Balto-Scandian craton, and it is worth considering whether top-to-the-east structures may be present among the mylonites in the HC. Although such top-to-the-*east* structures have not been observed in the top-to-the-*west* HC $\mathbf{S_3}$-mylonites near Sunnarviken, the existence of such contractional structures in the HC rocks in the N–NE subarea, or elsewhere in the study area, cannot be completely excluded. In outcrops where the HC mylonites lack shear sense indicators, both the top-to-the-west mylonites, and any potential top-to-the-east mylonites, would have to look quite similar, since it has not been observed that two types of mylonites, with distinctly different field apparences, are developed in the HC. This similarity in outcrop appearance would indicate that the potential top-to-the-east mylonites were formed under fairly similar metamorphic conditions as the top-to-the-west mylonites (although strain rate and other factors could have influenced the mylonite formation). However, as will be described later (Sect. 4.4), the metamorphic conditions were somewhat different during the top-E and top-W movements: $\mathbf{M_1}$-/$\mathbf{M_2}$-metamorphism of the top-E phase corresponds to the upper part of lower greenschist facies (chlorite grade), whereas the chlo+musc+semiductile qtz of the $\mathbf{S_3}$-mylonites indicate metamorphic conditions just *below* greenschist facies. Although it may be questioned whether mylonites of these metamorphic grades lie close enough to produce mylonites with quite similar field apparences, both mylonites would be of low grade. Therefore it cannot be exluded that mylonites of both types would look similar in the field. Hence, if for example the extensional $\mathbf{S_3}$-mylonites did not completely overprint the potential contractional mylonites, in the lower part of the HC, which would be the case if the extensional mylonites were developed in restricted zones only, the HC mylonites could contain top-to-the-east structures even though such structures were not detected in the present investigation.

4.3.7.4. COMPARISON OF TOP-TO-THE-WEST STRUCTURES OF THE STUDY AREA, WITH SIMILAR STRUCTURES IN THE HC ROCKS AT STANDAL

At the southernmost part of the Høydalsfjorden Complex, at a locality **6 km** SW of the southwestern boundary of the study area, reconnaissance work by the present author has revealed top-to-the-*west* kinematic indicators in HC-rocks exposed near the fjord immediately west of Standal, in the road-cut immediately north of the Standal fault. Shear bands appear to be particularly abundant at this locality. The brittle Standal fault separates the HC rocks to the north, from the Nordfjord–Sogn Detachment Zone (NSDZ) and the Western Gneiss Region (WGR) to the south. These top-to-the-west HC structures at Standal, which reflect movements similar to the above-mentioned top-to-the-west structures in the HC rocks between e.g. Vikafjell and

Sunnarviken, is interpreted to be a result of extension during the Devonian period. Also Krabbendahm & Dewey (1998) reported similar shear sense in these HC rocks near Standal, and they noted that no “true” mylonitic texture was developed, and that quartz behaved as porphyroclasts rather than in a ductile manner. Krabbendam & Dewey (1998) therefore suggested that the extension occurred in the semibrittle regime, and advocated that the temperature during deformation was much lower than that in the rocks south of the Standal Fault, i.e. in the NSDZ mylonites and the WGR. It appears that the properties of the sheared HC rock reported by Krabbendam & Dewey (1998) at Standal correspond fairly well to those found in the $\mathbf{S_3}$-mylonites described above from the HC rocks of the thesis area, although the quartz in the thesis area appears to be slightly more ductile than the ones at Standal. In terms of geographic distribution, it is not known whether the top-to-the-west structures are present only along the southern and northern contacts of the HC, i.e. along the Standal fault/NSDZ in the south and along the Sunnar fault/NSDZ in the north, or whether the structures occur elsewhere in the Høydalsfjorden Complex. It is suggested that the top-to-the-west structures are related to the Devonian extension. The textural and metamorphic properties of the rocks show that they were positioned at a shallow-crustal level.

4.3.7.5. COMPARISON OF TOP-TO-THE-WEST STRUCTURES OF THE STUDY AREA, WITH SIMILAR STRUCTURES AT STAVENESET

At the Staveneset peninsula to the west of the Kvamshesten Devonian Massif, Osmundsen & Andersen (1994) described abundant top-to-the-west kinematic indicators that were found in all tectonostratigraphic units above the Dalsfjord Suite; units which are, from bottom to top: the Høyvik Group, the Sunnfjord Melange, the Solund–Stavfjorden Ophiolite Complex, and the Stavenes Group. The structures were interpreted as related to Devonian extension. These top-to-the-west structures of the Stavenest peninsula appear to correspond to structures present in the HC rocks in the study area of the present thesis, as well as in the HC rocks near Standal.

4.3.7.6. CONCLUSIONS CONCERING THE HC $\mathbf{S_3}$-MYLONITES

Based on the present data from the study area, it must be concluded that the $\mathbf{S_3}$-mylonites of HC rocks display top-to-the-*west* shear sense, and that top-to-the-*east* shear sense indicators were not found. The HC $\mathbf{S_3}$-mylonites may have been formed contemporaneously with the *early* stages of the EG $\mathbf{S_2}$-mylonite fabrics. At a later stage, however, the HC $\mathbf{S_3}$-mylonites were cut by the Devonian unconformity, whereas the EG $\mathbf{S_2}$-mylonites continued to experience shear during formation of the HDM throughout the Middle Devonian period (**Table 4.1**). The fact that top-to-the-west structures are also found outside the study area (at the southwesternmost margin of the HC just north of the Standal fault) may indicate that the entire HC has experienced top-to-the-west shear.

4.3.7.7 POSSIBLE F_3-FOLDS

In several of the stereograms containing structural elements of the D_2-phase, a few fold axes with a roughly NE-SW trend are present (e.g. **Fig. 4.38a**, **4.39a**, **4.43a**), i.e. with a trend oriented nearly at right angle to the trend of the majority of the F_2-fold axes. As mentioned above, these deviating FAs are very rare. In the stereoplots, they constitute **19** of the **335** recorded FAs of the HC, or **5.7 %**. Likewise, a few axial plane poles and fold limb poles are located in the NW and SE parts of the stereoplots, (as opposed to the D_2-poles in the NE and SW quadrants), i.e. consistent with the deviating FAs. Although some of the D_3-structural elements are located not far from the “normal” D_2-pole groups, it cannot be excluded that the deviating structures are a result of a phase of F_3-folding (Sect. 4.3.7.2).

The NE-SW trend of the "deviating" HC F_3-FAs corresponds to a trend which has been thoroughly documented at the peninsula of Staveneset between Førdefjorden/ Stavfjorden and Dalsfjorden/Vilnesfjorden, i.e. west of the Kvamshesten Devonian Massif. There, the folds have been called F_3-folds, and have been interpreted as a result of Devonian /Post-Scandian extension-related top-to-the-west shear movements of the whole rock package, which gave the F_3-folds a roughly westward vergence (Osmundsen 1990; Osmundsen & Andersen 1992, 1994). In the Høydalsfjorden Complex, the presence of the NE-SW trending folds theoretically opens for the possibility that some of these folds also may have such a westward vergence. It is therefore theoretically possible that the HC contains a vague " F_3-phase" which may be related to the top-to-the-west S_3-mylonites in the HC, described above.

It has, however, not been possible in the *field* to see refolding of the F_2-folds by such potential "F_3-folds", nor have folds with a westward vergence been observed. Furthermore — as also mentioned above — the amount of "deviating" fold data is very limited compared to the major WNW-ESE F_2-trends, and the spreading of the F_2-fold axes around the mean trend may be fairly large. Based on the present data, it has therefore not been possible to actually prove that the "deviating" folds in the HC are related to the top-to-the-west movements in the area, and therefore it cannot be excluded that the folds are merely local deviations in the major F_2-trends. Deviations of this type have been seen to occur for example near the competent meta-gabbro bodies, and such bodies may have controlled the orientations of the F_2-fold axes locally. However, in the light of the presence of the top-to-the-west HC S_3-mylonites described above, as well as the numerous NE-SW trending folds described from Staveneset, it cannot be excluded that the deviating HC folds are in fact F_3-folds related to the top-to-the-west D_3-shear movement. The present investigations have, in all circumstances, shown that any such F_3-fold phase in the HC will be very vaguely developed compared to the Staveneset area.

4.3.8 POSSIBLE D_4-STRUCTURES

4.3.8.1 POSSIBLE F_4-FOLDS

The $\mathbf{S_3}$-mylonite fabric in the mountainside between Vikafjell and Sunnarviken (Sect. 4.3.7.1) has a very monotonous appearance, and the fabric is commonly without structural markers such as lithological banding, etc. In the field, the orientation of the mylonite fabric may thus be difficult to see. Nevertheless, the fabric generally appears to have a strike that trends E-W to NW-SE, and a dip towards N-NE or S-SW, where the amount of dip is variable. Systematic field measurements on the mylonite orientation have not been obtained, and neiter have possible systematic variations in orientation from outcrop to outcrop. It cannot be excluded that the apparent difference in dip of the fabric was formed during the $\mathbf{D_3}$-shear, in which case a phase of $\mathbf{F_4}$-folding would not be neseccary to explain the dip variation. However, the apparent variation in dip may also indicate that these rocks were in fact subjected to $\mathbf{D_4}$-contraction, producing $\mathbf{F_4}$-folds. Fold closures of such possible $\mathbf{F_4}$-folds have not been observed in the field, but this may be due to the limited degree of exposure. From the apparent strike orientation of the possible *limbs* to the $\mathbf{F_4}$-folds (i.e. the $\mathbf{S_3}$-mylonite fabric), it appears that the FAs of the possible $\mathbf{F_4}$-folds are trending E-W to NW-SE. The impression is, that the varying dips of the $\mathbf{S_3}$-mylonites were most likely caused by $\mathbf{F_4}$-folding.

As will be described in Sect. 5.5.2, the Håsteinen Devonian Massif has been folded about a WNW-ESE trending fold axis that is relatd to the massifwide $\mathbf{F_1}$-fold named the Osstrupen syncline. The *trend* of the HDM Osstrupen syncline is parallel to the clearly pre-Devonian HC $\mathbf{F_2}$-folds, and also parallel to the trend of the possible HC $\mathbf{F_4}$-folds. Since the possible $\mathbf{F_4}$-folds are folding top-to-the-west $\mathbf{S_3}$-mylonites that are cut by the sub-Devonian unconformity, it is possible that the Devonian sediments had been deposited by the time the possible $\mathbf{F_4}$-folding occurred. This would imply that the possible $\mathbf{F_4}$-folding of the HC occurred simultaneously with the $\mathbf{F_1}$-folding of the HDM (**Table 4.1**).

Apart from these possible $\mathbf{F_4}$-folds, however, it has not been possible to identify structures in the HC that could be related to the contractional phase that folded the Devonian ORS deposits. The apparent lack of deformation, in the HC, stemming from this contractional phase — i.e. deformation that would post-date the HC $\mathbf{F_2}$-folding — will be discussed later (Sect. 6.4.5.3).

4.4 METAMORPHIC DEVELOPMENT OF THE HØYDALSFJORDEN COMPLEX

Although not studied in detail, some aspects of the metamorphic grade related to the $\mathbf{D_1}$-, $\mathbf{D_2}$- and $\mathbf{D_3}$-deformation phases ($\mathbf{M_1}$, $\mathbf{M_2}$ and $\mathbf{M_3}$) will be briefly discussed.

4.4.1 M_1-METAMORPHISM

Metasediments: The $\mathbf{S_1}$-fabric in the meta-sediments of the HC contains the following metamorphic "index" minerals: *quartz, albite, white mica, chlorite, and epidote*. In addition, *biotite* was observed in a few instances, but this mineral generally appears to be absent in the study area. Small *garnets* (mostly **≤ 1 mm** in diametre) were observed at the Gravaneset peninsula (sub-area: W), in a grey-coloured meta-semipelite exposed within **a few metres** from the sea. In the Gravaneset area, meta-semipelitic rocks of this type generally form “layers” that alternate with “turbiditic” meta-greywacke, and the rocks have been intruded by numerous meta-gabbro bodies. At the garnet-bearing rock surface, the garnets were observed within an area measuring **a few square metres**. This is the only place in the HC where garnets have been observed in *ordinary* HC rock types. Due to this fact, the garnets are interpreted as a result of local effects, probably related to a high Mn-content in the sediments. This would allow growth of the garnet variant spessartine, which commonly grows below the garnet grade of the greenschist-facies (e.g. Yardley 1989).

Tiny garnets have also been observed at two other localities, in *not-ordinary* HC rocks: (i) in a black-coloured, high density, **50 x 20 cm** lense consisting of garnet-biotite-pyrite, located between Vikanipa and Sunnarviken (N-NE), where the garnets are **≤ 0.5 mm** in diametre; and (ii) in a **1 x 1 m** exposure of chert-like “ribbons” of interchanging quartz and chlorite, both with garnet, exposed in the road-cut near the top of the Langeneset peninsula. Here, the garnets are **≤ 0.1 mm** in diametre when coexisting with chlorite, and **0.01–0.02 mm** in diametre when associated with quartz. At least the Langeneset garnets are presumably of the Mn-rich type. The garnet-bearing rock-types at these two localities have not been observed elsewhere in the HC, and their very limited occurrence and exotic charachter open for the possibility that they are “clasts” with a melange origin, possibly with an inherited garnet assemblage.

Generally, the mineral paragenesis in the clastic HC rocks — as seen in more than **100 thin-sections —** indicate metamorphism corresponding to the chlorite grade, i.e. lower greenschist-facies conditions. The occurrences of biotite probably indicate that the metamorphism, in these few cases, just barely entered the biotite grade, apparently indicating that the general metamorphism of the HC corresponded to the *upper part* of the chlorite grade, i.e. upper part of the lower greenschist-facies. There is no sign of retrogression of a former higher metamorphic mineral paragenesis in the rocks. The minerals do not appear to have suffered retrogression, and the mineral association is therefore interpreted to have formed as a result of *prograde* metamorphism to the upper chlorite grade.

Meta-gabbro: In the meta-gabbro, the dominant minerals are strongly retrograded hornblende, sericitised plagioclase, actinolite, chlorite, epidote, and zoisite. The hornblende is extensively altered to actinolite, chlorite, epidote and white mica. Garnet has not been observed. According to Yardley (1989), development of garnet in meta-basites occurs at somewhat lower grades than in pelites, and any actinolite present in meta-basites will be altered to hornblende when garnet starts to grow. Biotite has been observed at one instance, but appears to be generally absent in the meta-gabbros of the study area. The presence of actinolite in the meta-gabbros thus supports the above assumption that the rare garnets in the meta-pelites are ***not*** of "garnet grade/upper greenschist facies", but spessartines of the chlorite grade/lower greenschist facies. In the HC gabbros, the existing mineral assemblage, including the retrograded hornblende — as well as the total absence of garnet and general absence of biotite — indicate that the rock experienced chlorite grade metamorphism, i.e. lower greenschist facies. The P/T-development in the meta-gabbros during $\mathbf{D_1}$ was prograde, and has led to retrograde mineral reactions of the original igneous gabbro minerals and growth of new metamorphic greenschist-facies minerals.

In summary, the meta-sediments and the meta-gabbros have thus experienced the same prograde development in pressure and temperature, producing a lower greenschist-facies $\mathbf{M_1}$-metamorphism.

4.4.2 M_2-METAMORPHISM

A distinct $\mathbf{M_2}$-metamorphic mineral assemblage cannot be distinguished. The $\mathbf{F_2}$-folds appear to merely fold and crenulate the $\mathbf{S_1}$-fabric, and no distinct $\mathbf{M_2}$-minerals, different from the $\mathbf{M_1}$-minerals, appear to grow in the crenulation fabric. Thin-section studies of hinges of $\mathbf{F_2}$-microfolds occationally reveal micas that are partly bent around the micro-fold hinges, i.e. being pre- $\mathbf{F_2}$ micas, and micas that partly form straight individuals with one end transgressing out of the microfold hinges, i.e. possibly indicating post-$\mathbf{F_2}$ growth of mica. This situation implies that the chlorite-grade P/T-conditions was maintained throughout the $\mathbf{F_2}$-folding. It can thus be assumed that also the $\mathbf{M_2}$-metamorphism corresponded to the upper chlorite grade, i.e. upper part of the lower greenschist-facies.

4.4.3 INTERPRETATION OF M1- AND M2-METAMORPHISM

The $\mathbf{M_1}$-metamorphism in the HC is *prograde* into the upper part of the chlorite-grade of the greenschist-facies, a grade which also continued during the $\mathbf{M_2}$-metamorphism. In terms of temperature, this metamorphism would roughly correspond to **350–400 °C**. If applying a thermal gradient of **30 °C/km**, the rocks experienced a burrial of roughly **12–13 km**. (A thermal gradient of **25 °C/km** would give **14–16 km** depth). In accordance with the interpretations of the $\mathbf{D_1}$- and $\mathbf{D_2}$-*structures* **(see Sect. 4.3.6),** this metamorphism is interpreted as related to the thrust emplacement of the HC onto the Balto-Scandian craton during the Scandian orogenic phase of the Caledonian orogeny.

The apparent lack of Devonian (post- $\mathbf{M_2}$-) retrogression of the chlorite grade mineral paragenesis in the HC is discussed later (Sect. 6.4.5.3).

Johnston et al. (2007b) briefly reported on metamorphism in the Høydalsfjorden Complex (which they named the Sunnarvik Group). The rocks were generally assigned to the greenschist facies, with a mineral assemblage of chlorite + muscovite + albite + quartz, +/- biotite, +/- epidote. This degree of metamorphism is consistent with the conclusions of the present thesis. However, Johnston et al. (2007b) also carried out thermobarometry on a rock from Stavøya island, located in the westernmost part of the Høydalsfjorden Complex. The analyses yielded P = **9.1 +/- 1.3 kbar** and T = **442 +/- 71 °C**, which was estimated to indicate burrial to **~30 km** depth. The sampled rock were termed to be a *gneiss*. Comparison was made with the Upper Allochthon rocks on Bremangerlandet, which were reported to display greenschist–low amphibolite facies fabrics. The metamorphic grade of the sampled rock obviously differs strongly from what is common in other parts the Høydalsfjorden Complex (this work). If these P/T results are correct, the author of the present thesis will propose that the sampled rock probably belongs to a different unit than the rest of the Høydalsfjorden Complex, or alternatively that the sampled rock is part of an olistolith in a melange. More thermobarometric data is needed to clarify this issue.

4.4.4 M_3-METAMORPHISM

The $\mathbf{M_3}$-metamorphism can be observed in the $\mathbf{S_3}$-mylonites. In the field, the penetrative mylonitisation in the zones leave no traces of the protolith, but the dominating grey-coloured variant is interpreted as a meta-sediment. As briefly mentioned in Sect. 4.3.7.1, thin-section studies show that the mylonites contain extensive growth of chlorite and white mica. Frequently, the texture is “bi-modal”, i.e. separated into a matrix and “porphyroclasts”. The matrix is then dominated by the chlorite and white mica, and also by black bands of cryptocrystalline minerals that prevent transmittance of normal light under the microscope; whereas the “porphyroclasts” consists of quartz, and also of feldspar and epidote. The feldspar displays brittle behaviour. It is sometimes seen that during the shear movements, “porphyroclastic” grains of quartz have disintegrated into smaller pieces, a process which has in some cases been initiated by brittle cracking of the quartz, although the brittle cracking has immediately afterwards been followed by recrystallisation that produced numerous tiny quartz sub-grains. Non-“porphyroclastic” quarts grains are generally distributed in the texture, and also this quartz shows extensive development of tiny sub-grains. The combination of quartz “porhyroclasts” that show signs of cracking, and the contemporaneous extensive development of tiny subgrains, indicate a temperature of **~ 300 °C** (Passchier & Trouw 1996) during the $\mathbf{D_3}$-shear process. With a thermal gradient of **30 °C/km**, this implies a burrial of roughly **10 km**. Accordingly, the degree of metamorphism correponds to a level just *below* the greenschist facies.

In the Eikefjord Group (EG), the metamorphism during the **D_2**-mylonitisation of the EG rocks has been shown to be ***retrograde*** (Sect. 3.5.2), and has been interpreted as related to the post-Scandian/Devonian extensional movements along the *Nordfjord–Sogn Detachment Zone* (Sect. 3.4.2), bringing the mylonitic rocks from lower amphibolite-facies to lower greenschist-facies. In the HC rocks, the **M_3**-metamorphism reflects low-grade metamorphic conditions in the *Upper Plate*, i.e. *above* the Nordfjord–Sogn Detachment Zone. At the Staveneset peninsula between Førdefjorden/Stavfjorden and Dalsfjorden/Vilnesfjorden, Osmundsen & Andersen (1994) described greenschist-facies metamorphism related to top-to-the-west shear deformation there, and such top-to-the-west shear fabrics and realated metamorphism have also been observed in the HC, in the form of the **D_3**-features. The **M_3**-metamorphism in the HC appears to be of lower grade than the middle to lower greenschist-facies metamorphism of the EG **S_2**-mylonites. This shows that the HC **S_3**-shear occurred at a somewhat shallower crustal level than the **S_2**-shear in the EG, which is exactly what would be expected from their tectonostratigraphic positions. The HC **S_3**-mylonites and the EG **S_2**-mylonites were partly developed in the same top-to-the-west shear event. The *early*-stages of the NSDZ movements (forming EG **S_2**-mylonites), which are reflected by the amphibolite facies remnants in the EG **S_2**-mylonites, might have been contemporaneous with the HC **S_3**-mylonites (**Table 4.1**). It is, however, likely that the *later*-stages of the EG **S_2**-shear, which are reflected by the lower to middle greenshist facies minerals, postdate the HC **S_3**-shear, since the NSDZ (i.e. EG **S_2**-mylonites) was active for a long time period during the formation of the Håsteinen Devonain Massif, whereas the HC **S_3**-mylonites are cut by the Devonian unconformity below the Håsteinen deposits.

4.4.5 ABSENCE OF M_4-METAMORPHISM

It has *not* been observed that the **S_3**-mylonite fabric, with its **M_3**-metamorphic minerals, has been recrystallised by an **M_4**-metamorphic phase that could be related to the possible **F_4**-folds. If the possible **F_4**-folds were formed during the contractional phase that also folded the Håsteinen Devonian Massif (HDM) into the Osstrupen **F_1**-syncline, it appears that this contraction did not impose recrystallisation of the HC **S_3**-/**M_3**-fabrics. The same situation applies to the **M_1**-/ **M_2**-minerals in the **S_1**-/ **S_2**-fabrics, which also show no signs of a possible **M_4**-related recrystalisation. However, as will be discussed later, the folding of the HDM produced an axial planar cleavage in sandstones in the westernmost parts of the Devonian massif, implying that temperatures were high enough for crystal plastic processes to take place during the folding of the HDM. The same temperature was obviously present in the immediately underlying HC rocks, yet apparently did not produce any **M_4**-related recrystallisation in these rocks.

4.5 SUMMARY AND CONCLUSIONS FOR THE HØYDALSFJORDEN COMPLEX

The Høydalsfjorden Complex consist of meta-sediments and gabbroic meta-intrusives that formed in a marginal basin located between a subduction-related volcanic island-arc and a continental margin. In addition, the Kvangagjelet syenite, which is located at the southern margin of the Håsteinen Devonian Massif is possibly an integral part of the Høydalsfjorden Complex, although this is at present uncertain. No marker lithology exists in the HC rocks of study area, and it is thus not possible to reveal the three dimensional architecture or stratigraphic relationships of the sedimentary rocks in the complex.

Based on geochemistry and general geology, Furnes et al. (1990) included the rocks of the present Høydalsfjorden Complex in their original Heggøy Formation, which was defined from the Staveneset area. There, the rocks of the Heggøy Formation were interpreted as the primary *sedimentary cover sequence to the Solund–Stavfjorden Ophiolite Complex (S–SOC)*, which formed in a marginal basin. It is possible that certain portions of the HC rocks form a part of the cover sequence to the S–SOC. However, other parts apparently do not, justifying the new name Høydalsfjord Complex for the rocks.

The rocks of the HC may also represent a melange being equvivalent to the *Sunnfjord Melange*, which has been defined from the Staveneset area and interpreted as an obduction melange related to the obduction of the Solund–Stavfjorden Ophiolite Complex during the Caledonian Orogeny (Andersen et al. 1990; Alsaker & Furnes 1994). The Kvangagjelet syenite could then theoretically be a mega-olistolith in such a melange, although this is at present uncertain.

The HC-rocks may also correspond to the *Kalvåg Melange*, which is located at the Frøya and Bremangerlandet islands W to WNW of the Hornelen Devonian Massif. This melange has been interpreted as a *sensu stricto* sedimentary olistostromal melange that formed by slumping and sliding on the submarin slope of a volcanic-arc island bordering the marginal basin in which the S–SOC formed.

The D_0-deformation is related to soft-sedimentary processes. In the areas containing interbedded "turbiditic" meta-greywacke and meta-semipelite, the structures may be "chaotic". These structures may have been formed by slumping and sliding processes. The large numbers of meta-gabbro bodies, interpreted to have intruded at the oceanic stage, may also have contributed to the possible syn-sedimentary deformation during the intrusion processes.

The D_1-deformation has produced the $\mathbf{S_1}$-cleavage/-foliation, which is the dominant tectonic fabric in the HC rocks. The $\mathbf{S_1}$-fabric is seen to be essentially parallel to $\mathbf{S_0}$ wherever the bedding is present. The $\mathbf{S_0}$-/$\mathbf{S_1}$-parallelism may have been accomplished in four ways; through: (i) isoclinal $\mathbf{F_1}$-folding of $\mathbf{S_0}$; (ii) predominantly bedding-parallel non-coaxial deformation; (iii) predominantly bedding-orthogonal coaxial contraction; and (iv) combi-nations of (ii) and (iii). The $\mathbf{M_1}$-metamorphic conditions that existed during formation of the $\mathbf{S_1}$-fabric corresponded to upper part of the chlorite grade, i.e. upper part of the lower greenschist-facies, at **350–400 °C**.

<u>The D_2-deformation</u> led to folding of the **S_1**-cleavage into **F_2**-folds that have FAs with a WNW-ESE trend and low to moderate plunge in both directions. The **F_2**-folds are the dominating fold structures in the HC rocks. At the outcrops, the amplitudes and wavelengths vary, but are commonly in the order of **5–10 m**. Normally, these meter-scale folds do not have a penetrative axial planar **S_2**-cleavage. In stead, **S_2** is defined by kink bands or, locally, spaced crenulation cleavage, that may vary in consentration from locality to locality. The kink bands are usually developed in conjugate sets. The **M_2**-metamorphic conditions during the **D_2**-event was the same as for **D_1**, which means upper part of the chlorite grade, i.e. upper part of lower greenschist facies, at **350–400 °C**. The **F_2**-folds are cut by the Devonian unconformity, and thus predate deposition of the Devonian sediments.

From the island of Stavøya in the westernmost part of the Høydalsfjorden Complex (HC), thermobarometry performed by Johnston et al. (2007b) yielded **9.1 +/- 1.3 kbar**, **442 +/- 71 °C,** from a rock which they termed to be a gneiss. They estimated the P/T to reflect a depth of **~30 km**, and the metamorphism to be lowest amphibolite facies. The P/T data are much higher than the lower greenschist facies recorded elsewhere in the HC, and if these P/T estimates are correct, it is proposed (this work) that the sampled rock belong to a different unit, or alternatively that it be an olistolith in a melange. More data is needed to resolve the matter.

The HC **F_2**-folds and the EG **F_3**-folds have the same WNW-ESE trends, but the folds are *not* genetically related. HC **F_2**-folds are related to the contractional Scandian eastward obduction and nappe transport of the HC rocks on the Baltic craton. The HC **F_2**-folds are furthermore cut by the Devonian unconformity and therefore pre-date the deposition of the Devonian sediments. In the EG, **F_3**-folds probably formed during the extensional top-to-the-west movements on the Nordfjord–Sogn Detachment Zone, possibly contemporaneously with the folding of the Devonian ORS deposits, and the EG **F_3**-folds are thus later than the HC **F_2**-folds.

The **D_1**- and **D_2**-features were formed during obduction of the outboard terranes onto the Balto-Scandian margin during late Silurian times. These processes resulted from the Scandian phase of the Caledonian Orogeny, which was the main phase of continent-continent collision between Baltica and Laurentia.

<u>The D_3-deformation</u> is represented by **S_3**-mylonites that show top-to-the-west sense of shear. These HC mylonites are cut by the sub-Devoniain angular unconformity, showing that they pre-date deposition of the Håsteinen Devonian Massif. The top-to-the-west sense of shear suggests that the mylonites are related to post-Scandian extensional movements. Indications of top-to-the-*east* sense of shear have not been observed. A few recordings of folds that trend NE-SW, i.e. orthogonal to the majority of the **F_2**-folds, may possibly represent **F_3**-folds. Folds with this NE-SW trend have been reported from the Staveneset peninsula west of the Kvamshesten Devonian Massif (Osmundsen 1990; Osmundsen & Andersen 1992, 1994), where the folds have been interpreted as a result of top-to-the-west shear movements affecting the whole Caledonian nappe pile. In the Høydalsfjord study area, **M_3**-metamorphic conditions during formation of the **S_3**-mylonites corresponded to a temperature of **~300 °C**, i.e. below the greenschist facies.

Southwest of the study area, at the ***road-cut*** located by the sea-side just west of <u>Standal</u> and immediately north of the brittle Standal Fault, the HC-rocks display penetrative top-to-the-west shear fabrics that are consistent with the movements seen in the **S_3**-mylonites of the study area.

The HC $\mathbf{S_3}$-mylonites in the study area may have been formed simultaneously with the ***early stages*** of the EG $\mathbf{S_2}$-mylonites. The HC $\mathbf{S_3}$-mylonites are, however, cut by the Devonian unconformity, showing that the movements on the HC $\mathbf{S_3}$-mylonites terminated well before deposition of the Devonian sediments. Following this termination of movements on the HC $\mathbf{S_3}$-mylonites, the EG-$\mathbf{S_2}$-mylonites (NSDZ) had a prolonged shear history throughout the Middle Devonian period, with extensional movements that facilitated westward movement of the Upper Plate, leading to formation and deposition of the Middle Devonian Håsteinen Devonian basin unconformably on top of the HC $\mathbf{S_3}$-mylonites.

The existence of a $\boldsymbol{D_4}$*-deformation* phase is uncertain. This deformation, if present, are represented by $\mathbf{F_4}$-folds, and might be responsible for variations in dip of the mylonitic $\mathbf{S_3}$-fabrics. The strike of the $\mathbf{S_3}$-fabrics appears to have a trend around WNW-ESE, and the $\mathbf{S_3}$-fabric would then constitute the fold limbs to $\mathbf{F_4}$-folds, which themselves would also trend WNW-ESE. The possible $\mathbf{F_4}$-folds might have formed during the Devonian deformational phase that also led to folding of the Håsteinen Devonian deposits, into the WNW-ESE trending $\mathbf{F_1}$-Osstrupen syncline. It has, however, not been possible to find conclusive evidence showing that this deformational phase affected the subjacent HC-rocks.

The trend of the three generations of FAs of the HC, notably (i) the $\mathbf{F_2}$-folds; (ii) the possible HC $\mathbf{F_4}$-folds; and (iii) the HDM Osstrupen $\mathbf{F_1}$-fold, are parallel.

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Chapter 5

<u>CHAPTER 5 HÅSTEINEN DEVONIAN MASSIF (HDM)</u>

<u>5.1 INTRODUCTION</u>

General

The present chapter deals with the geology of the Håsteinen Devonian Massif (HDM), which occupies the central parts of the study area **(Fig. 5.1** and **Plate 1)**. Irgens & Hiortdahl (1864) introduced the name "Håsteinen" for this massif, a name whivh was taken from the highest mountain peaks in the southernmost part of the area displaying Devonian rocks (**Fig. 1.3**). The HDM is assumed to be of Middle Devonian age by correlation with other west Norwegian Devonian massifs which have been dated by plant and fish fossils (see e.g. Steel et al. 1985). The chapter will present data on the <u>*western*</u> half of the HDM, i.e the area situated to the west of a line running N-S through the lake Vassetvatnet. The size of the Devonian area studied is about **25 km^2**.

Purpose

The purpose of this chapter is to describe (1) the Devonian rocks; (2) the sub-Devonian unconformity; (3) the inliers of Høydalsfjorden Complex rocks within the Devonian massif; and (4) the structural and metamorphic development of the HDM.

The HDM has not been subjected to a separate investigation since the reconnaissance paper of Kolderup (1925) which contains simple petrographic descriptions and mentions the primary unconformity in the western parts. Seranne's (1988) investigation of some structural features of the HDM was very limited, as it formed a minor part of a large study on all the Devonian massifs in western Norway. Since the geology of the HDM is therefore largely unknown, one object of the present investigation has been to elucidate all major aspects of the geology of the massif.

The Devonian geological development of western Norway has been the focus of renewed interest in the last decades due to the development of the "Detachment model" (Sect. 2.8). In this situation, it has been unfortunate that the *structural* (and metamorphic) geology of the HDM has been largely unknown, and a major object of the present broad study has therefore been to describe details of the *tectono-metamorphic* history of the HDM.

Content

The chapter starts with a description of the sedimentology of the HDM, and also briefly presents the intrusive hydrothermal epidote-quartz dykes (Sect. 5.2). This is followed by a presentation of the unconformities below the HDM (Sect. 5.3). Thereafter the intra-Devonian inliers of Høydalsfjorden Complex rocks are described and interpreted (Sect. 5.4). This sub-chapter is followed by a presentation of the structural and metamorphic development of the HDM, including features such as folding, axial planar cleavage, stratigraphic thickness, faulting and the three dimensional architecture of the massif (Sect. 5.5). The final sub-chapter contains a summary and the main conclusions (Sect. 5.6).

Terminology and abbreviations

As already indicated, the name "Håsteinen Devonian Massif" will be abbreviated **"HDM"**. In this work, the HDM, strictly speaking, means the *western half* of the Håsteinen Devonian Massif, i.e the area located within the present study area. When explicitly stated in the text, the "HDM" may, however, also mean the *whole* massif, i.e. including the eastern parts. The term **"massif"** is used for the *present* outcrop of Devonian rocks, whilst the term **"basin"** is used for the *original* (and presumably much larger) depositional area. The term **"Håsteinen Group"** (and the associated formations) refers to the sedimentological / stratigraphical organisation of the sedimentary rocks, whilst the term "Håsteinen Devonian Massif" is used for the Devonian rocks as a unit, although covering exactly the same rocks as the "Håsteinen Group". The term "Håsteinen Group" has also been used by Bryhni & Lutro (1991a, 1991b). The term **"substrate"** is used according to Torsvik et al. (1988) to denote all the various rocks underlying the HDM. The substrate rocks forming the westward-oriented wedge seen to go "into" the Devonian rocks on Teigafjell is called the **"Teigafjellet Substrate Wedge"** and is abbreviated **"TSW"** (see Sect. 5.4.3).

Highlights of the HDM

The following is a rewiew of interesting features characterizing the HDM, that will be discussed in the following sub-chapters.

- More than **99 %** of the massif consists of conglomerates and breccias, which are generally monotonous and structureless. Small sandstone bodies are found only in the far west, just above the basal contact.
- Small lenses of planar laminated sandstone, with orientation parallel to the conglomerate beds, have revealed the bedding orientation in the conglomerates and breccias troughout he HDM.
- Two different clast populations are present; meta-sedimentary clasts in the northern half and meta-igneous clasts in the southern half of the massif; with the two clasts types interfingering in the west.
- Primary unconformable Devonian/substrate contacts are abundant along the HDM margin, implying that the basin is sitting *in sit*u on its immediate substrate.

- Three outcrops of substrate inliers occur as windows within the HDM, implying that the cover of Devonian sediments is generally thin.
- Bedding is everywhere steep (**55–75°**), defining a massif-wide syncline plunging steeply ESE. The northern limb is oriented with a NE strike and a steep SE dip, and the southern limb with a SE strike and a steep NE dip. The axial plane is sub-vertical and the axial trace has a WNW-ESE trend.
- Small-scale parasitic folds with axial planar cleavage are developed in a sandstone body in the western part of the HDM. The HDM was the first Devonian deposit in Norway where folds with axial planar cleavage were reported (Vetti 1988,1989); but such structures have later been described by Sturt & Braathen (2001) from the Solund Devonian Massif.
- Cumulative stratigraphical thickness of the studied part of the HDM is **5.8 km,** established along the WNW-ESE trending axial trace (**—** and **11.2 km** for the whole massif), whilst the "basin" measures only **7.2 km** horizontally along this trace (**— 14 km** for the whole massif). The "basin" thus has the largest cumulative stratigraphical thicness compared to its length, of all the Devonian "basins" of western Norway. The vertical thickness is only somewhat more than **1 km**.
- Metamorphism in the sediments corresponds to the anchizone/prehnite-pumpellyite facies or lowermost greenschist facies.
- Faults cutting the HDM have only minor displacements.
- The present-day substrate-"topography" (underneath the Devonian sediments) displays a relief of at least **500 m**; as can be seen from the map, where the position of the unconformable Devonian/substrate contact varies greatly in altitude.
- In the eastern parts of the study area, the everywhere *steeply-dipping* bedding, which is part of the basin-wide Osstrupen syncline, stands with a high angle towards a sub-Devonian unconformable contact that has a *semihorizontal* envelope surface. Unfolding and back-rotation of the Devonian beds to the subhorizontal orientation they had at the time of deposition, thus have severe consequences for the original orientation of the underlying unconformable Devonian/substrate contct, since this contact would be rotated to a steep “unrealistic” dip. Consequently, a new model is needed for the formation and deformation of the HDM, a model which must be able to explain this extraordinary geometry of “steep-bedding-on-horizontal-contact”. Such a model and its regional implications are discussd in Ch. 6.

5.2 ROCK DESCRIPTIONS

5.2.1 GENERAL

Section 5.2 presents a description of the rocks in the HDM. Emphasis is on the sedimentary Devonian rocks, although a very brief description of locally occurring epidote-quartz dykes and quartz-veins is included. Since the Devonian conglomerates and minor sandstones are practically undeformed at outcrop-scale, the rock descriptions of Sect. 5.2 mainly deals with the *sedimentology* of these rocks.

The section starts with a presentation of the Håsteinen Group including descriptions of conglomerate formations; an apparently regolithic breccia; sandstone units; possible palaeocurrent directions; and possible source rocks (Sect. 5.2.2). After the subsequent description of intrusions/veins in the Håsteinen Group (Sect. 5.2.3), a summary is given (Sect. 5.2.4).

Apart from the general rock descriptions, the chapter focuses on two subjects; (1) the sand lenses revealing bedding orientation in the otherwise massive conglomeratic HDM, and (2) the interpretation of the depositional environment of the conglomerates – which may provide information on a possible inclination of bedding at the time of deposition. These two points deal with the orientation of bedding, a subject which will be of particular significance, since the orientation of bedding (see **Plate 1**) forms the basis for the discussion in Sect. 5.5 on the structual development and the three dimensional architecture of the HDM.

5.2.2 THE HÅSTEINEN GROUP

General

The sedimentology of the HDM has not been investigated since Kolderup (1925) presented the first reconnaissance study of the basin and reported on its conglomeratic nature and on the clast-types. The following presentation of sedimentological aspects consentrates on the main features and presents an overall picture.

The Håsteinen Group

The Håsteinen Group, as defined here, embraces all the Devonian sediments in the study area. The sediments contain two main facies; coarse conglomerates/breccias, which totally dominate the HDM (>99 % by area), and small units of sandstone which are located in the western and northwestern parts. The conglomerates are divided into two formations; the Vikafjell Formation and the Blåfjell Formation. In addition, a breccia which apparently represents a sub-Devonian regolith, is present at Litledokka, in the east. Six sandstone units are present; being named (1) Mannen, (2) "Galmannsskåra", (3) Stegabruna, (4) Novene, (5) Vikaholmen and (6) Gravaneset sandstone units (**Fig. 5.2** and **Plate 1**).

On the preliminary 1:50.000 geological survey maps "Eikefjord" and "Naustdal", Bryhni & Lutro (1991a, 1991b, 2000a, 2000b) also used the designation "Håsteinen Group" for the Devonian sediments in the study area, and they also included the Devonian sediments located to the *east* of the present study area in their group. Bryhni & Lutro (1991a, 1991b, 2000a, 2000b) furthermore used the term "Svardal Formation" for *all* different types of conglomerates and breccias in the area. However, due to the recognition, in the present work, of *several* conglomerate formations, it is recommended that the designation "Svardal Formation" be abandoned.

Organisation of the section

Section 5.2 presents the conglomerates and the sandstones in two separate subsections. The conglomerates are presented first, although the sandstone units are generally located stratigraphically below the conglomerates (see Sect. 5.4). The reason for this order of treatment is that the sandstone occurrences are very small and therefore less important in a massif-scale structural context, compared to the conglomerates. The interpretation of the depositional environment for the conglomerates are presented after the description (Sect. 5.2.2.1). Thereafter, the six sandstone units are treated, followed by a presentation of the depositional environment (Sect. 5.2.2.2).

Terminology

The lithostratigraphical term "formation" is used without definition of type-sections, and should therefore be considered as informal naming. The formations are, however, mappable units, have well defined boundaries, etc., and therefore otherwise satisfy the formal requirements for having "formation" status (Nystuen 1989). The sandstone bodies are relatively small, and are therefore termed "units".

5.2.2.1 CONGLOMERATES

The conglomerates are divided into two formations; the Vikafjell Formation and the Blåfjell Formation, and this subdivision is based on a clear bimodal distribution of clast types in the basin. A "boundary-zone" separating the two clast-type areas can be drawn from Blåfjell in the west-southwest and eastwards via Osen and along the Osen meta-psammite inlier (**Fig. 5.2**, **Plate 1**, **Plate 2**). The clasts to the south of this zone and, hence, to the south of the Osen meta-psammite, have been derived from mainly magmatic rocks (syenitic-monzonitic-gabbroic in composition). The clasts to the north of the zone are mainly of meta-psammitic composition. In the areas to the east of Osen, the change from one clast type to the other occurs relatively abruptly, i.e. within **a couple of hundred metres** in the map plane. In the western parts, the transition is very gradual and occurs over a width of more than **800 m**.

It is worth noting that the meta-psammitic clasts correspond to the lithologies in the Høgdene and Osen inliers, and that the syenitic/monzonitic clasts correspond to the Kvangagjelet syenite (see Sect. 5.4)

THE VIKAFJELL FORMATION

General

The formation is defined by the conglomerates consisting of a high proportion of meta-psammitic clasts, usually more than **80-90 %** (see below). The name of this formation is taken from the mountain ridge of Vikafjell (**Fig. 1.2** and **Plate 1**) located in the northern part of the HDM, and the name is proposed for the first time in the present work. As lichen is usually absent in these higher regions, the formation is very well exposed. It constitutes the whole area to the north of the clast-type boundary-zone (**Fig. 5.2** and **Plate 1**), except for the marginal parts in the western and northwestern areas where the sandstone units of Vikaholmen, Gravaneset, Mannen, "Galmannsskåra" and Stegabruna are located. This means that the areas from around Blåfjellnipa and westwards/northwest-wards consists of the Vikafjell Formation.

At the lower boundary, the formation rests either directly on (1) the substrate (Sect. 5.3), (2) the Mannen sandstone unit (see below), (3) the "Galmannsskåra" sandstone unit (see below), (4) the Gravaneset sandstone unit (see below), or (5) the intra-Devonian Høgdene substrate inlier (Sect. 4.2.1.4 and 5.4). The transitions to the underlying sandstone units are gradual. The upper boundary is unknown, as the formation continues outside the study area towards the E.

Description

In terms of facies, the formation consists of pebble, cobble, boulder and block conglomerates and breccias. The clast population consists of more than **90 %**, usually foliated meta-psammites of whitish colour and arkosic to quartzitic composition (**Fig. 5.3**), together with minor amounts of other clasts (see below). The bedding-orthogonal stratigraphic thickness of the formation is abo ut **3.5 km** within the study area. This is a minimum estimate, since the formation continues east of the study area. It should be noted that the layers that constitute this steeply-dipping "stratigraphy" was not deposited vertically on top of each other (see discussion in Sect. 6.3.2); since the present stratigraphy is the result of syn-sedimentary, successive, tectonic roll-over rotation and subsidence of the layers in a half-graben setting during westward movement of the basin on the subjacant extensional Nordfjord–Sogn Detachment Zone. The clasts in the Vikafjell Formation were derived from lithologies that corresponds to the rocks presently occurring in the local substrate defined by the Høydalsfjorden Complex, although the meta-psammitic clasts are substantially over-represented compared to their abundance in the substrate. Nevertheless, in the substrate, large areas containing the same type of meta-psammites are present, especially between Kalvik and Sandvika (Kildal 1970), in addition to the intra-Devonian Høgdene meta-psammite inlier, the Osen meta-psammite inlier, and the meta-psammite on Fjellsenden (see further discussion on basin-fill source rocks in Sect. 5.2.2.4). The conglomerates and breccias are clast-supported (**Fig. 5.3**), polymict, and contain the following clast types: meta-psammites (normally constituting **80-90 %** of the clasts, see **Fig. 5.3**), and in addition minor amounts of meta-semipelite (qz-fsp-mica-schist) (**Fig. 5.4**), meta-greywacke, meta-pelite, meta-gabbro, meta-basalt, greenschist and amphibolite. The clast size varies from pebble (**Fig. 5.5**) to block (**Fig. 5.6**), and the largest clast found – located to the north of the summit of Mannen (at UTM 0789 3122) – has a cross-sectional diameter of **1.7 x 1.2 m**. The conglomerates and breccias are very poorly sorted (**Fig. 5.4** and **5.6**), and all clast sizes are normally present. Occasionally, clearly bimodal deposits are present. This is the case, for instance, at the Stigen Devonian Spur, where a bed of boulder-sized clasts now contains sandstone in the interstitual places (**Fig. 5.7**). The deposit is interpreted as a boulder-containing debris flow that has lost its original finer material to the subjacent strata, and where the resulting open spaces have later been filled with sand from above. This kind of deposit has been called "sieve deposit", following Hooke (1967).

The degree of rounding varies from angular, via subangular (**Fig. 5.6**) and subrounded (**Fig. 5.3**), to rounded, although most clasts are subangular to subrounded. The clast-shapes are usually tabular (elongated) for foliated clasts (**Fig. 5.4**) and more equidimensional for less foliated clasts. The clast orientation is usually random and chaotic (**Fig. 5.6**), but bedding parallel orientation, or imbrication have occationally been observed.

The thickenss of the conglomerate beds tend to be difficult to estimate when the average clast size in the various beds is not significantly different. When observable, the thickness of the beds varies, and may be up to at least **5 m**. The basal contacts tend to be difficult to recognise when the beds are structureless and have uniform clast sizes, but when a sand lens is present on top of the underlying bed, or the clast size is different in the beds, it is possible to see the basal contact (**Fig. 5.8**). The basal contacts are presumably somewhat erosive, at least towards the sand lenses, although channels have not been observed. Grading within the flows are occasionally present (**Fig 5.8**), both normal and revers, but frequently the flows are massive and ungraded (**Fig. 5.6**).

A clearly separate matrix is normally not present in the conglomerate beds, as they are usually polymodal and not bimodal in particle size. The intra-clast structures are very variable. In places, the undeformed arkosic meta-psammitic clasts may show nice primary cross-bedding internally (**Fig. 5.9**). It should be noted that the Sandviken meta-psammite, in the substrate to the west-northwest of the study area, shows well preserved primary sedimentary structures (**Fig. 4.8**), indicating that such features were present in the source areas (Sect. 5.2.2.4) during deposition of the Devonian sediments. Other clasts contain a foliation which is folded, showing that the source area had undergone at least two phases of deformation prior to the deposition of the Devonian sediments. The amount and style of deformation in these clasts corresponds well to that of the local substrate of the HDM. On the Stigen Devonian Spur, a clast was observed to show signs of disintegration, probably as a result of rough transport (**Fig. 5.10**).

THE BLÅFJELL FORMATION

General

The Blåfjell Formation is defined by a very high portion of magmatic clast lithologies, usually more than **80-90 %**. The Vikafjell and Blåfjell Formations are lateral eqivalents, since they grade into each other in the west. Along the southern margin of the Osen meta-psammite inlier, the transitional zone between the two formations displays a mixture of the two clast types (**Fig. 5.11**). On weathered surface the meta-psammitic clasts stand out due to their larger weathering resistance (**Fig. 5.11**). The name of the Blåfjell formation is taken from the mountain ridge of Blåfjell located in the southwestern part of the HDM (**Fig. 1.3** and **Plate 1**), and is proposed for the first time in the present work. Even though the rock assigned to this formation is generally, in the elevated areas, free of both vegetation and Quaternary cover, an extensive cover of lichen (enhanced by the relatively high content of mafic minerals in many of the clasts) is typical, making it very difficult to find good exposures (**Fig. 5.12**). The formation constitutes the area to the south of the clast-type boundary zone (**Fig. 5.2**), i.e. approximately the southern half of the HDM. From Fjellsenden in the east, the formation is present all the way to the areas around Blåfjellnipa in the west.

At its lower boundary, the Blåfjell formation rests with a primary unconformity on the substrate, or on the Kvangagjelet syenite. The upper boundary is again unknown, as also this formation continues outside the study area towards the E.

Description

In terms of facies, the formation consists of coarse conglomerates and breccias. The clast population largely consists of magmatic rocks of syenitic-monzonitic-gabbroic compositions (**Fig. 5.13**). The clasts in the Blåfjell formation do not reflect the abundant substrate rocks of meta-greywacke, meta-semipelite, meta-gabbro and meta-psammite around the HDM. Instead the plutonic syenite-monzonite clasts are similar to the syenite/monzonite that constitute the "exotic" Kvangagjelet syenite (Sect. 4.2.2.4 and 5.4). The gabbroic clasts could theoretically be of the same type as the meta-gabbros present in the meta-greywackes of the substrate, but the absence of other clast types from the substrate suggests that the gabbro clasts are not derived from this kind of source. The fact that the clasts of the Blåfjell formation are almost exclusively derived from magmatic lithologies, whereas the Vikafjell Formation, on the other hand, contains essentially whitish meta-psammitic clasts, indicates that the formations had different source areas (see Sect. 5.2.2.4 for further discussion on source rocks). The bedding-orthogonal stratigraphic thickness of the formation is about **4 km** within the study area. This is a minimum estimate, since the formation continues towards the east. It should again be noted that the layers that constitute this steeply-dipping "stratigraphy" was not deposited vertically on top of each other (see discussion in Sect. 6.3.2); since the present stratigraphy is the result of syn-sedimentary, successive, tectonic roll-over rotation and subsidence of the layers in a half-graben setting during westward basin movement on the subjacant extensional Nordfjord–Sogn Detachment Zone.

The conglomerates are polymict (**Fig. 5.13**), with clasts of mainly syenite, monzonite and gabbro, together with small amounts of greenstone, granite, amphibolite, and quartzitic to arkosic meta-psammites. The conglomerates are usually clast supported (**Fig. 5.13**), and the clast sizes show a complete range of sizes, where the largest clasts have a cross-sectional diametre of at least **0.5 m**. They are generally poorly sorted (**Fig. 5.13**), and a distinguishable matrix is normally not present in the conglomerate beds, i.e., the clast size distribution is generally not bimodal. The degree of rounding is hard to observe due to the extensive cover of lichen, but the clasts seem to be "sub-rounded to sub-angular". The clast-shapes are highly variable, but since the magmatic clasts have no foliation that yields a preferred splitting direction, tabular clasts are less abundant than in the Vikafjell Formation. Clast orientation is usually random, but indications of bedding-parallel orientation or imbrication have been observed. Due to the extensive cover of lichon, it has not been possible to see whether the beds are graded. It is, however, reasonable to assume that the situation regarding grading is the same as for the Vikafjell Formation, i.e. that normal graded, reverse graded, and ungraded beds are present. Bed thickness and basal contacts of various beds are difficult to observe due to lichen, but again it is assumed that the the situation is similar to what was found in the Vikafjell Formation. Intra-clast structures may be discerned in the magmatic clasts which usually contain a massive magmatic texture with no well defined tectonic fabric.

DEPOSITIONAL ENVIRONMENT OF THE VIKAFJELL AND BLÅFJELL FORMATIONS

Although the exposure of the Blåfjell Formation is limited by an extensive cover of lichen, the sedimentology of the Blåfjell and Vikafjell Formations appears to be very similar. The two formations are thus treated together in the interpretation of the depositional environment.

The conglomerates are interpreted as debris flow deposits. As shown above, the conglomerates defining one flow (bed) are usually coarse, polymodal, clast supported, poorly to very poorly sorted, and lack any pervasive stratification. The flows have clast fabrics which are largely random, i.e. there is a lack of imbrication and other types of preferred orientation of longest clast dimensions, and the flows are usually ungraded. Outsized boulders and blocks are occationally present. There also appears to be a lack of any significant erosion between flows, and the flows have a sheet-like form. According to Nemec & Steel (1984), these features are typical of subaerial debris flow deposits on alluvial fans. Similar deposits in the marginal fans of the Hornelen Devonian Massif were interpreted by Gloppen & Steel (1981) and Larsen & Steel (1978) as debris flow deposits in an alluvial fan environment. Also the Kvamshesten massif contais such deposits (Osmundsen et al. 2000). The coarse and almost structureless conglomerates and breccias in the Håsteinen Group are therefore interpreted to have been deposited in an alluval fan environment by subaerial debris-flows.

The most coarse and structureless conglomeratic to breccia-like deposits may have originated from scree/rock fall and grain avalanche deposition as described from the Hornelen Devonian Massif (Larsen & Steel 1978). In alluvial fan settings, streamflood and sheetflood deposits, etc. may occur (Larsen & Steel 1978), and such deposits may be present in the western parts of the Håsteinen Group. The debris flow deposits are, however, predominant.

SANDSTONE LENSES IN THE VIKAFJELL AND BLÅFJELL FORMATIONS

General

Within the coarse conglomerate deposits of the Vikafjell and Blåfjell Formations, very small and discontinuous layers of granular sandstones and pure sandstones are occasionally present. The layers are up to **a couple of metres** long and **some decimetres** thick (see below). The layers are for short, termed sandstone *"lenses"* due to their limited latteral extent. The lenses are important because they have made it possible to study the orientation of bedding in the otherwise almost structureless Vikafjell and Blåfjell Formations, and thus to establish the structural geometry of the folded/rotated bedding in the massif (Sect. 5.5). Two types of lenses appear to be present. The most common lens type is located *between* the individual debris flows and is termed "inter-flow" lenses. The other type appears to be located *within* the individual debris flows and is called an "intra-flow" lenses.

Description of inter-flow lenses

The lenses are scattered throughout the HDM in an unsystematic way and are present as cappings on the flows.

The internal development of the lenses varies. The lenses may consist of essentially pure sandstones (locally graded), gravelly sandstones, or interlayered gravels and sandstones (**Fig. 5.14**, **5.15** and **5.16**). The lenses are sometimes graded with fine massive conglomerate at the bottom giving way to granular sandstone and proper sandstone towards the top (**Fig. 5.16**). Also the more pure sandstone lenses are occasionally graded. Coarsening upward has been observed in places (**Fig. 5.14b**), but most lenses show upward-fining (**Fig. 5.16**). It is assumed that this is normal grading, suggesting right way-up. However, since reverse grading may also be present, the lenses as such are unreliable as way-up indicators. The sandstone lenses are sometimes structureless and massive**,** or partly massive (**Fig. 5.15**), and may be planar laminated (**Fig. 5.15** and **5.16**). Cross-stratification has not been observed in the lenses, and the planar stratification within the lenses is, when present, parallel to the elongated shape of the lenses themselves.

The external shape of the lenses are usually elongated, typically of the order of **1–2 m** in length and **10–30 cm** in thickness (**Fig. 5.14** and **5.16**). The largest lense observed was located **50 m** to the east-southeast of the summit of Nonsnipa (near Osen). The planar laminated "sandy" part of this "lens", which could be observed in “3D view”, was **2 m** long, **3 m** deep and **15 cm** thick, but in addition there were **15–20 cm** of planar stratified gravel and gravely sandstone grading into the proper sandstone. Generally, the lenses are always thin compared to the underlying debris flow bed.

It is possible that the capping sand lenses originally formed somewhat more extensive sheets or layers which were partly removed by the next flow, i.e. that the present lenses are erosional remnants, but clear evidence for this has not been found.

Interpretation of inter-flow lenses

Similar lenses and layers (cappings) on top of sub-aerial debris flows have been described from the marginal alluvial fans of the Hornelen Devonian Massif by Larsen & Steel (1978) and Gloppen & Steel (1981). They suggest that the capping is the product of waning-stage flow and winnowing of the upper part of the debris flow by waterflow following the termination of the debris flow episode (Gloppen & Steel 1981; Jahns 1947; Johnson 1970).

Nemec & Steel (1984) have investigated the presence of analogous upward-fining gravely-sandy to sandy cappings with or without an erosive base and signs of stratification, in areas of sub-aerial debris flow deposits. They interpreted the layers to be a result of turbulent fluidal flow or heavily sediment-laden stream flow following a debris flow. The sandy deposits of such a flow are interpreted as a result of waning traction currents. The sandy layers are frequently planar stratified, and the planar stratification are reported to be either flat lying, i.e parallel to the overall bedding surface of the sheet-like debris flows, or shallowly inclined (possibly in the range of **0–10°**) compared to the overall debris flow bedding.

The sandstone lenses in the HDM are accordingly interpreted as the effects of minor fluvial processes acting on the top of just settled debris-flow deposits, forming discontinuous cappings on the flows. The water was supplied at the end of a debris flow event and the competence of the flow was only large enough to carry/support sand and gravel.

Intra-flow lenses

Occasionally, thin (**5–10 cm**) discontinuous sand layers appear to be present *within* the debris flow beds. Such sand layers may be interpreted as the result of rapidly surging debris flows (Nemec & Steel 1984). The layers are generally "flat-lying" and parallel to the debris flow beds and, hence, also reflect the overall bedding of the debris flows.

Discussion: the use of sandstone lenses as bedding indicators

Most bedding measurements in the field are based on sandstone lenses, so it is important to show that these are also representative for the overall bedding of the debris flows. Debris flow deposits generally form sheetlike beds, and the basal erosion is limited and insignificant (Nemec & Steel 1984). Therefore, the interfaces between the beds are generally fairly good bedding markers representing the surface of the alluval fan at the time of deposition. Whenever the debris flow beds in the Håsteinen Group could be separated, the orientation of internal planar lamination in the lenses, and the orientation of the longest dimension of the lenses themselves, were compared with the general orientation of the debris flow beds. The *strike* of internal lamination, as well as of the longest dimensions of the lenses themselves, generally appeared to be parallel to the strike of the interfaces between adjacent beds of debris flows when such interfaces could be seen, thus showing that the lenses provide relatively good estimates of the strike orientation of bedding. Also in the *dip* direction the planar lamination in the lenses, as well as the lens shape itself, appeared to be parallel to the flow interfaces. Although a general parallelism appears to exist between the sandstone lenses and the sheet-like debris flow beds, variations are, of course, present. This variation must be at least **+/- 10°**. In areas with a large number of bedding measurements (e.g. the Nonsnova-Teigafjell area (**Plate 1**), these variations are displayed. The bedding recordings, are, however, remarkably consistent, and the variations are thus averaged out. This suggests that the planar stratification within the lenses and the outer shape of the lenses reflect the overall orientation of the debris flows quite well. The planar stratifications in the lenses are thus good bedding indicators, and the orientations of the stratification have been recorded as representative of bedding.

Generally at the stage of depositon, the surface of alluvial fans, and hence the debris flow beds themselves, have a shallow primary inclination. This was studied by Anderson & Cross (2001) at the northern margin of the Hornelen deposits, where the authors compared the angle between alluvial fan slopes and the ajacent down-stream braidplane deposits, of which the latter were deposited close to horizontal. The analyses showed that the alluval fans had a primary basinward slope of **4°** at the time of deposition. The issue of primary alluval fan orientation is discussed later in relation to the folding/rotation of the Devonian bedding (Sect. 5.5.2.1). The orientation of bedding, and the three dimensional architecture of the HDM, is treated in Sect. 5.5. Mean values of bedding orientations are also plotted on the main map (**Plate 1**).

THE LITLEDOKKA BRECCIA

At Litledokka (**Plate 1**) in the easternmost part of the study area, a breccia is present which is very different from the conglomerates/breccias of the Vikafjell and Blåfjell Formations. The clasts of the Litledokka Breccia are completely angular (**Fig. 5.17**), as opposed to the subangular/subrounded clasts of the Vikafjell and Blåfjell Formations, indicating that the sediments have been subjected to very limited transport. The clasts are foliated meta-psammitic and meta-semipelitic rocks (or qz-fsp-mica-schists) of precisely the same type as the local substrate surrounding the HDM. Bedding or stratification appear to be absent, and the deposit therefore appears to be massive. The clasts are mainly of pebble to cobble size, but boulders also occur. The deposit is clast-supported, poorly sorted, polymict, polymodal, ungraded and without particular clast orientations. The contact towards the rocks of the Blåfjell Formation can be observed **400 m** to the east-southeast of the summit of Storedokka mountain (UTM 1308 2728) (**Fig. 5.18**). The contact is of primary depositional type and is sharply defined.

The Litledokka Breccia is interpreted as a combination of a regolith and scree deposits that lie *in situ* or have been subjected to only very little transport. The deposit has later been burried by the debris flows of the Blåfjell Formation. On the main map (**Plate 1**), the contact between the Litledokka Breccia and the Blåfjell Formation is denoted "Primary sedimentary contact; possibly conformable". The contact is "conformable" in the sense that the breccia was probably unconsolidated/unlithified before the deposition of the Blåfjell Formation, although there may, of course, have been a limited time gap (hiatus) between the deposition of the two units.

5.2.2.2 SANDSTONE UNITS

The Håsteinen Group contains 6 sandstone units which shall be briefly described in the following. These are the (1) Mannen, (2) "Galmannsskåra", (3) Stegabruna, (4) Novene, (5) Vikaholmen and (6) Gravaneset sandstone units. These names are proposed for the first time in the present work. The units are first described individually, and thereafter interpreted collectively in terms of transport mechanism and depositional environment.

MANNEN SANDSTONE UNIT

The unit is named from the mountain-top of Mannen on the northwestern part of Vikafjell. The unit is situated along the northwestern margin of the HDM just to the north and north-northeast of the summit of Mannen (**Appendix A: map sheet no. 5b** and **5c**, and **Plate 1**). The length of the sandstone body is about **50 m**, and the thickness usually **5–10 m**, with maximum thickness in the range of **20 m**. The lower boundary is the depositional angular unconformity towards the substrate, and the upper boundary is a gradual transition into the conglomerates of the Vikafjell Formation. The unit contains the following facies: massive sandstone, bedded massive sandstone, and planar laminated sandstone. In addition, the sandstones alternate with layers of pebbly conglomerates.

At two localities, about **600 and 800 m** to the west of the Mannen sandstone unit, discontinuous Devonian sandstone layers with a thickness of **0.5–1.0 m** are present on top of the substrate along the Devonian/substrate contact. The transition between the sandstone layers and the Vikafjell Formation is sharp. The layers are considered too small to be given status as separate sandstone units, and will not be further discussed.

STEGABRUNA SANDSTONE UNIT

The name of the unit is taken from the 1:5.000 economic map and comes from the area just to the west of the Høgdene area. This sandstone unit occurs on the steep montain side below the Intra-Devonian Høgdene meta-psammite inlier (**Appendix A: map sheets no. 6a** and **6b**, and **Plate 1**), and is present from a point about **300 m** to the east of the Høgdene Lake and westwards to at least **a couple of hundred metres** to the west of "Galmannsskåra", giving it a length of about **1.8 km**. The thickness of the unit varies between **a few metres** and **20 m**. The unit is not continuously exposed latterally, and although the unit is assumed not to continue all the way to the Mannen unit **a couple of hundred metres** further west, the possibility that it does continue cannot be completely ruled out. In the mountainside, the "lower boundary" of the unit is a depositional angular unconformity towards the substrate, and the "upper boundary" is a similar depositional unconformity towards the meta-psammites of the Høgdene inlier. The unit contains the following facies: massive sandstone, bedded massive sandstone, planar laminated sandstone, and layers of pebbly conglomerate are present throughout.

"GALMANNSSKÅRA" SANDSTONE UNIT

The name is from the northern part of the central areas of Vikafjell. "Galmannsskåra" is a name that is “constructed” by the author of this thesis, after the nearby situated "Galmannstufta" (from the economic map), to denote the cleft going up the entire northern steep mountainside to the Vikafjell plateau (**Plate 1**). The sandstone unit can be studied at the upper part of the cleft of "Galmannsskåra", at a little shelf/plateau just to the east of the fault that is located in "Galmannsskåra" (**Fig. 5.19**), and the unit is there lying on top of the Høgdene meta-psammite inlier. The unit is at least **50-100 m** long. The east-northeastern termination of the unit cannot be observed as the unit continues over the cliff edge on top of a vertical cliff at the end of the plateau (**Fig. 5.19**). Yet, sandstone has been observed further east, at the top of the river section about **125 m** to the east of the little plateau, and this sandstone may possibly represent an eastward continuation of the unit. The sandstones do, however, definitly terminate before the top of the eastern part of the Høgdene meta-psammite. The sandstone has a total thickness of about **5 m** at the "Galmannsskåra" locality (**Fig. 5.19**). The lower boundary is a depositional angular unconformity towards the meta-psammites of the Høgdene inlier (**Fig. 5.20**), and the upper boundary is a transitional or abrupt change into the conglomerates of the Vikafjell formation (**Fig. 5.21**).

In terms of facies, the sediments in the unit consist of fluvially sorted sandstones, with layers of pebbly conglomerates (**Fig. 5.22**) that frequently wedge out. Frequently, the sandstones are massive to structureless internally (**Fig. 5.23**), but the beds themselves may be clearly defined (**Fig. 5.23**). The sandstones may also be planar laminated. Planar laminated sandstones with small-scale planar cross-stratification (**Fig. 5.24**), and ripple cross-laminated sandstones (**Fig. 5.25**), are present, but are rare. Both these cross-stratified structures suggest local transport towards the NE after "back rotation" of bedding to sub-horizontal position. Alternating layers of sandstone and pebbly conglomerate are present towards the transition to the overlying conglomerates of the Vikafjell Formation (**Fig. 5.23**). To the east of the locality on the plateau area, in a vertical cliff at the top of the river section mentioned above, thin layers of fine sandstone/siltstone are present. At this locality, current ripples (seen from underneath in a vertical cliff) are present in sandstones (**Fig. 5.26**). The ripples indicate that the local transport after "back rotation" was towards the WSW. The presence of both NE and WSW transport direction — i.e. practically opposite — in the same area shows that the local variations in transport directions have been very large, and these flow structures has therefore not been used to deduce general transport directions in the basin.

GRAVANESET SANDSTONE UNIT

The name of this unit is taken from Gravaneset promontory at the eastern end of Høydalsfjorden (**Plate 1**). The sandstone unit itself is located near the tip of the promontory at the westernmost areas of the HDM, and is partly cut by the road section at Gravaneset (**Fig. 5.27**, see also **Plate 4** and **Plate 5**). The length of the unit is at least **120 m** in the strike direction, and it may continue even further beneath Quaternary deposits towards the south. The thickness of the unit is at least **100 m**, and it is thus the thickest sandstone unit in the Håsteinen Group. The lower contact is defined by the depositional unconformity towards the substrate of Høydalsfjord Complex rocks (**Fig. 5.27**), and the upper contact is a gradual transition into the debris flow conglomerates of the Vikafjell Formation.

In terms of facies, the Gravaneset sandstone unit consists mainly of fluvially sorted, interlayered sandstones and pebbly conglomerates. The sandstones are of a type similar to the deposits in the coarse parts of the "Galmannsskåra" unit. At the base of the Gravaneset unit, near the unconformity in the southern part of the road section, a **5 m** thick conglomeratic layer is present (**Fig. 5.28**). This particular conglomerate layer appears to wedge out within **10 m** towards the north. The rest of the unit contains bedded sandstone, massive sandstone, planar laminated sandstone, and small-scale cross-stratified sandstone (**Fig. 5.29**), with frequently ocurring layers of pebbly conglomerates of the same type as in the "Galmannsskåra" unit. The cross-stratification indicates that the local transport direction was towards the WSW after "back rotation" of bedding to the sub-horizontal orientation, but due to the possibility of large local variations of transport directions in these sandstones (see the "Galmannsskåra unit"), general transport directions for the basin will not be inferred from this structure.

VIKAHOLMEN SANDSTONE UNIT

The name is taken from the small island of Vikaholmen where the sandstone is present on the southeastern and eastern part of the island (**Plate 4**). The unit is about **20–30 m** long in the strike direction and possibly about **5 m** thick. The lower boundary is located below the sea level, and the upper boundary is a sharp transition to debris flow conglomerates. The facies in the unit are massive structureless sandstone, massive bedded sandstone and small-scale cross-stratified sandstone. Single pebbles may be present in the planar laminated sandstones.

NOVENE SANDSTONE UNIT

The name of this unit is taken from the ridge of Novene at the west-southwestern margin of the HDM, where the sandstone is present precisely on the crest of the ridge (**Plate 1**). The length of the sandstone along strike is about **5 m**, and the thickness is at least 1-**2 m**, and possibly **4 m**, although this is uncertain due to lack of exposure. The lower boundary of the unit is not exposed, although a depositional unconformity between conglomerates and the substrate is exposed near by. The upper boundary is defined by an abrupt sedimentary transition into the conglomerates of the "clast-boundary-zone" (= mixed pebble zone) present between the Vikafjell and Blåfjell Formations. In terms of facies, the unit contains more massive, structureless sandstone as well as planar laminated sandstone (**Fig. 5.30a**) that is sharply cut by the massive debris flow conglomerates on top (**Fig. 5.30a)**. The debris flows appear to have eroded channels into the sandstones, and conglomerate has later filled the eroded depressions (**Fig. 5.30b**).

DEPOSITIONAL ENVIRONMENT OF THE SANDSTONE UNITS

All the sandstone units are located directly on top of the substrate rocks of the Høydalsfjorden Complex, either with a depositional unconformable contact, or with a tectonically modified depositional contact. The mutual stratigraphical position of the units is uncertain, since no stratigraphical markers exist in the areas of massive debris flow deposits between them. The sandstone units do, however, appear to represent the lowest stratigraphic level in the Håsteinen Group. An overview of the distributions of the various facies in the sandstones are shown in **Table 5.1**.

All the units (except the very small Novene unit) are characterised by a variable degree of alternating layers of sandstones and conglomerates as illustrated from the "Galmannsskåra" unit (**Fig. 5.22** and **5.23**) (**Table 5.1**). The overall layering must have been sub-horizontal at the time of deposition. Nemec & Steel (1984) have interpreted this type of fluvial facies-association as typical of braided stream environments. All the units contain massive structureless sandstones, bedded massive sandstones and planar laminated sandstones (except the small Novene unit which appears to contain *only* planar laminated sandstone and more massive sandstone) in close association with the intercalated layers of conglomerates. Also these facies types must therefore be interpreted as related to the same braided river environment. Although rare, the occurrences of cross-stratification in the sandstones of the "Galmannsskåra" and Gravaneset units may also indicate braided stream environments.

Although the maturity varies, the conglomerates associated with the sandstones are generally texturally immature (low degree of sorting, rounding, etc). According to Nemec & Steel (1984) this immaturity suggests deposition from ephemeral (flash) flooding in alluvial fan stream environments, although separation from debris flows may at times be difficult.

The alluvial-fan braided-stream environment may encompass deposits of "primary fan channel", "distributary braided stream", and "sheetflood/streamflood" types (Nemec & Steel 1984). A discussion of these sub-environments are, however, beyond the scope of the present study, and should be undertaken in future sedimentological investigations.

When the facies associations in the sandstone units are considered in the light of the coarse alluvial fan conglomerates deposited on top of the sandstones, it is possible that the sandstones are deposited in a distal-fan environment. The presence of conglomerates on top of the sandstone units may be interpreted as the result of the out-building of more proximal fans on top of the more distal sandstones as the fans developed.

5.2.2.3 PALAEOCURRENT DIRECTIONS: DISCUSSION

An investigation of palaeocurrent directions was not a part of the present study, but some aspects of the subject will be discussed.

As mentioned above, presence of small-scale cross-stratification (current ripples) in the sandstone unit of "Galmannsskåra" (**Fig. 5.24**, **5.25** and **5.26**) show local transport directions both towards the NE and the WSW (after "back rotation" of bedding to "original" sub-horizontal position) — i.e. opposite diretions — and similar structures in the Gravaneset unit (**Fig. 5.29**) show local transport towards the WSW (after "back rotation"). However, since the transport direction in these sandstones may, due to channeling, etc., show extreme variations on a local scale, as illustrated above from the "Galmannsskåra" unit, these local directions will not be used to deduce the overall transport directions in the Håsteinen Devonian Massif. Instead, the transport directions will be discussed in terms of facies and clast distributions, and also in the light of the possibility that a basin controlling fault was located along the southern margin of the HDM:

(1) Location of sandstone units in the W may indicate E to W transport: The sandstone units are only found in the western (WSW to WNW) part of the massif at the lowest stratigraphical level present, suggesting that these parts were the most distal parts at the time of sandstone deposition. If the sanstone units represent basin-axis sandstones, this may suggest palaeocurrents directed towards the W, WSW or WNW. Such a westward transport direction would be compatible to the axially transported sandstones in the Hornelen Devonian Massif (Steel et al. 1985) and the Kvamshesten Devonian Massif (Osmundsen et al. 1998). If, however, the sandstones are distal parts of fans, other directions cannot be excluded, such as from the south margin towards the N, and from the north margin towards the S, which would also be compatible to the marginal fan sandstones (and conglomerates) in the Hornelen Devonian Massif (Steel et al. 1985) and the Kvamshesten Devonian Massif (Osmundsen et al. 1998).

(2) Different transport directions in north and south: Conglomerate beds of the “meta-sedimentary clast” Vikafjell Formation in the north and the “meta-igneous clast” Blåfjell Formation in the south interfingers in the west. This implies that any particular stratigraphic "level" continues from one formation into the other, and each "level" consists of two distinctly different clast types in north and south. Any one such stratigraphical "level" cannot have been fed from only *one* source alone, since this would imply a systematic down-fan or lateral change in clast population of the *single* fan during deposition, with repetition of this process for every new flow/bed. In addition the size of such a single, basin-wide fan would be too large to be geologically realistic. It may thus be concluded that the Håsteinen Group has been fed from two completely different source areas. This, in turn, suggests that the transport directions for the deposition of the north and south parts were different. It is therefore possible that the Vikafjell Formation was fed essentially from the north and the Blåfjell Formation from the south, in accordance with the situation at the south and north margins of the Hornelen and Kvamshesten massifs. These Håsteinen conglomerate formations are situated stratigraphically above the sandstone units.

(3) South of the HDM, the presence of the NSDZ and the superimposed brittle Standal Fault may imply transport S → N from the southern margin: The situation at the southern margin of the HDM is very similar to the situation at the southern margin of the Hornelen Massif. A segment of the mylonitic Nordfjord–Sogn Detachment Zone (see Sect. 2.8) is present, with a brittle, steeply N-ward-dipping Mesozoic(?) fault (Standal Fault in the Håsteinen area) following within/along and cutting the zone. The southern marginal fans of

the Hornelen massif show transport towards the north, i.e. there has been a source area to the south. The original Devonian brittle faults in the Upper Plate, that controlled the Devionian basins, must have been located just north and south of the present margins of the Hornelen basin, since the marginal fans must have been only slightly larger when juging from the fan sedimentology. The Håsteinen Massif formed a result of Devonian brittle faulting in the Upper Plate, probably occurring at a time when the Upper Plate was positioned further east. The later Mesozoic(?) faults are located quite near and parallel to the traces of the original Devonian faults. The later Mesozoic(?) faulting has produced the present juxtaposition of the HDM/HC rocks against the detachment zone mylonites. It is possible that the presence of the mylonites to the south of the HDM, in combination with the Standal Fault, may suggest the presence of a source area in the south at the time and place of Håsteinen deposition. However, one must remember that a potential original source area in the south would – after the Mesozoic extensional faulting – be left behind further east.

In summary, a sediment transport from the southern margin towards N, from the eastern margin towards W, and the northern margin towards S may be suggested. This would agree with the transport directions reported from the other Devonian massifs along the coast of Western Norway (Steel et al. 1985; Osmundsen et al 1998). It should be noted that the *present* margin of the HDM aginst the Høydalsfjorden Complex was probably not the original basin margin, as the basin presumably extended further westwards on the Høydalsfjorden Complex.

5.2.2.4 SOURCE ROCKS FOR THE HÅSTEINEN GROUP

As outlined in Sect. 5.2.2.1, the northern part of the HDM consists essentially of foliated and massive/unfoliated meta-psammitic clasts (Vikafjell Formation), and the southern part essentially of apparently unfoliated syenitic, monzonitic and gabbroic clasts (Blåfjell Formation). Since the palaeo-transport directions are uncertain (Sect. 5.2.2.3), the exact geographical "palaeo-" directions to the original source areas for the clasts are not known, although a general transport direction along the basin axis from the E towards the W, and locally at the northern, eastern and southern margins from the N, the E or the S respectively, may be assumed.

The internal metamorphic and structural state of individual clasts reflect the degree of metamorphism and the nature of deformation in the source rocks prior to deposition, and may thus help to identify the source area. The metamorphism in the meta-psammitic clasts appears to correspond to greenschist-facies. Deformed clasts show one major foliation (S_1) which has been folded (F_2). Mylonitic clasts have not been observed. The metamorphic state of the magmatic clasts are more unclear, but also they appear to have suffered greenschist-facies metamorphism. The magmatic clasts appear to be practically undeformed.

According to the "Detachment model" (outlined in Sect. 2.8), the areas between Sognefjorden and Nordfjord may be assigned either to an Upper Plate or a Lower Plate, separated by the mylonitic Nordfjord-Sogn Detachment Zone (see **Figs. 2.3**, **2.7** and **2.8**). In the Håsteinen area, the Nordfjord-Sogn Detachment Zone (NSDZ) follows along the southern margin of the HDM (e.g. **Fig. 2.8**), and also along the northern margin, where it incorporates all rocks of the Eikefjord Group to the north of the HDM (e.g. **Fig. 2.3**). The Devonian rocks of the Håsteinen Massif are situated on the greenschist- facies Upper Plate rocks (of the substrate/Høydalsfjorden Complex). The "eclogite-facies" Lower Plate is present to the east and south. The present separation into a greenschist-facies Upper Plate, generally in the west, and an "eclogitic" Lower Plate, generally in the east, is a result of the extraordinary uplift of the Lower Plate east of the NSDZ, during Devonian crustal-scale extension.

At the time of ORS deposition, the presently exposed "Lower" and "Upper Plate" areas contained essentially the same types of rocks, i.e. the "greenschist-facies" rocks. Devonian basins formed on top of these greenschist-facies rocks, and surrounding rocks of similar type acted as source-rocks for the ORS sediments. Later, tectono-extensional and erosional processes have removed the original "greenschist facies" source rocks from the Lower Plate areas. The Lower Plate experienced extreme uplift, leading to tectonic and erosional removal of the greenschist-facies rocks, eventually revealing the uplifted high-grade "eclogitic" terrane (**Fig. 2.7**), which are now present to the south of the detachment zone going along the southern margin of the HDM, as well as to the east of the HDM (**Fig. 2.3**). The original ***types*** of rocks that were present in the source area for the Håsteinen Group must therefore now be sought among the greenschist-facies rocks present in the substrate below the Devonian deposits, i.e. the Upper Plate. The original ***geographical position*** of the source areas, however, was most likely further east in the areas now occupied by the high-grade Lower Plate rocks.

Source-rocks for Vikafjell formation clasts:

The clasts in the Vikafjell Formation are completely dominated by meta-psammitic rocks. In addition, minor amounts of meta-semipelite, meta-greywacke, meta-pelite, meta-gabbro, meta-basalt, greenschist and amphibolite are present. The meta-psammites in the Grøneheia area (**Fig. 2.5**) to the north of the village of Eikefjord are extensively mylonitised, and are present in an area that has been interpreted as part of the Nordfjord–Sogn Detachment Zone (see Ch. 3). However, since mylonitic clasts appear to be absent in the HDM, these meta-psammitic rocks (and the mylonitic detachment zone rocks in general) have not been the source for the meta-psammitic or the other clasts in the Vikafjell Formation. However, the meta-psammitic rocks located in the Høydalsfjorden Complex to the west of the HDM, e.g. at Sandvika and the islands of Alvora and Stavøya (**Fig. 2.4**), are of a type that is analogous to the clasts in the HDM with respect to lithology, structural development and metamorphism. The substrate also contains rocks corresponding to the other clast types in the formation, with exception of the amphibolites which have not been observed in the substrate. Hence, rocks that are analogous to the meta-psammites and other rocks in the substrate may thus have been present in the original source area for the clasts.

Source-rocks for Blåfjell Formation clasts:

The clasts in the Blåfjell Formation are dominated by syenite, monzonite and gabbro. In addition, minor amounts of greenstone, granite, amphibolite and meta-psammite are present. The meta-gabbroic rocks now present in the substrate to the HDM are closely associated with meta-greywackes, etc., and since meta-sedimentary clasts appear to be rare in the Blåfjell Formation, it is unlikely that the gabbro clasts were derived from this type of rock association. On the other hand, the small amounts of greenstone, meta-psammite and possibly granite could have come from these types of rocks. Instead, the predominance of magmatic clasts makes a plutonic complex more likely as the main source. Such a complex could be of either Caledonian or Precambrian age. Caledonian candidates that could represent analogs to the syenitic/monzonitic and gabbroic clasts are possibly present. The syenitic/monzonitic clasts are identical to the Kvangagjelet syenite at the southern margin of the HDM, on which the conglomerates rest with a depositional unconformable contact. If the Kvangagjelet syenite is an integral part of the substrate, and not a mega landslide in the HDM, this body would indicate that syenitic/monzonitic clasts could have originated from rocks analogous to the present substrate. If the Kvangagjelet syenite is, however, a landslide in the HDM, it would be an open question whether the syenite clasts could be connected to the substrate rocks. Intrusives like the Bremanger Granodiorite (dated to **440 +/- 5 Ma**, U-Pb/zircon, Hansen et al. 2002) and the Gåsøy Gabbro (dated to **443 +/- 4 Ma**, U-Pb/zircon, Pedersen pers. com in Hansen et. al 2002), both situated to the west of the Hornelen Devonian Massif, indicate that Caledonian plutonic rocks were possibly also present in the Håsteinen source area. Precambrian candidates for the different *types* of source rocks are present in the Dalsfjord Suite located between Førdefjorden/Stavfjorden and Dalsfjorden/Vilnesfjorden (Sect. 2.5). This kind of suite contains a wide range of little deformed plutonic rocks, such as mangerite syenites, gabbros and anorthosites (Kolderup 1921). (Dalsfjord Suite intrusive age dated to **1634 +/- 3 Ma**, Corfu & Andersen 2002). The presence of the Dalsfjord Suite immediately below the Kvamshesten Devonian basin shows that such plutonic rocks were present in subaerial position at the time of the deposition of the Devonian sediments, and therefore could have acted as source rocks. Radiometric dating of the igneous clasts in the Blåfjell Formation should be carried out in the future to test these hypotheses.

Source-rocks for the Litledokka Breccia clasts:

The Litledokka Breccia contains clasts of foliated meta-psammitic and meta-semipelitic rocks (or qz-fsp-mica-schists). The clasts are identical to the local substrate around the HDM, and the source for these clasts must have been situated very near the breccia. It is thus likely that the topographical relief that led to the formation of the breccia was located near by. The relief was possibly formed by the original Devonian fault that controlled the deposition of the breccia and the rest of the Håsteinen Group. In accordance with the situations at the southern margins of the Hornelen and Kvamshesten Massifs, such a Devonian basin-controlling fault may have been present just to the south of the brittle Mesozoic(?) Standalen Fault, which follows along, and cuts through the mylonites of the Nordfjord-Sogn Detachment Zone south of the HDM (**Plate 1**).

In conclusion, the source rocks for the clasts in the HDM generally occupied a shallow crustal position (greenschist-facies) during the Caledonian orogenic cycle. Such rocks are now present only in the Upper Plate lithologies to the west of the Nordfjord–Sogn Detachment Zone. The essentially meta-psammitic clasts of the Vikafjell Formation could well have been derived from meta-psammitic rocks of the same type as those presently situated in the substrate to the HDM. The magmatic clasts of the Blåfjell Formation are either derived from a Caledonian plutonic complex, or from Precambrian plutonic rocks similar to the ones in the Dalsfjord Suite immediatly below the Kvamshesten Devonian Massif. The qz-fsp-mica-schist clasts in the Litledokka Breccia are derived locally.

5.2.3 "INTRUSIONS"/VEINS IN THE HÅSTEINEN GROUP

Epidote-quartz dykes

The only “intrusive” dykes in the HDM are quartz-epidote dykes and veins. These are present at a number of localities, but particularly good examples are found on the island of Parisholmen located at the western end of the fjord Osstrupen (**Plate 1**), and in the Gravaneset sandstone unit (at localities just to the east of the electric power lines). The colour of the dykes is light to darker yellowish green, and the colour is a characteristic feature. At Parisholmen, the main dyke is **10–15 cm** thick (**Fig. 5.31**), and oriented with a strike/dip of **124/54 SSW**. Xenoliths of vein-quartz are present. At Gravaneset, two types of dykes occur. The first type is uncommon, as it has a darker green colour than usual. The other type is the typical strongly yellowgreen-coloured type, here with a thickness up to **7 cm**. These veins may form irregular patterns (**Fig. 5.32**). The dykes cut the sub-Devonian depositional angular unconformity (see Sect. 5.3.2.2).

The time of intrusion is uncertain, and the intrusive activity might have occurred over a long time span. It is likely that the intrusions occurred during the Devonian tectonothermal event which also deformed the Devonian rocks. The epidote component has most likely been derived from the gabbroic/basaltic rocks in the substrate.

Quartz -veins

Quartz-veins are rare in the HDM and have only been found at a few localities. Examples are found at Straumsnes **a few metres** to the east of the road, and at the Gravaneset sandstone unit, on top of the cliff, to the east of the electric power lines. At Straumsnes, at least two localities are present. At a locality **50 m** to the east of the road, the veins are normally **20–60 cm** long and **a few millimetres** thick (**Fig. 5.33**). The quartz-veins are here fairly parallel, with an average orientation of **060/60 NW**, and, hence, do not appear to form net-veins (**Fig. 5.33**). At a locality **a few metres** to the east of the road cut, quartz-veins with an average orientation of **070/70 NW** cut one set of epidote-quartz-veins, although it may not be concluded from this that the quartz-veins post-date all epidote-quartz-veins. At Gravaneset, the single quartz-vein is up to **5 cm** thick. The quartz veins cut Devonian clasts, showing that the rock was well lithified at the time of vein implacement.

Although the exact time of formation of the veins is not known, it is possible that the quartz veins formed partly as a result of local pressure solution between Devonian clasts during the Devonian tectonothermal event. Although conclusive evidence of such pressure solution has not been observed in the field, the process might have occurred on a smaller scale. It is also possible that the veins were developed from hydrous solutions saturated with dissolved quartz, which circulated in the rock during the Devonian tectonometamorphic event. Local dilation may have facilitated the emplacement of the veins.

Veins in other Devonain massifs in western Norway

In the three Devonian massifs of Hornelen, Kvamshesten and Solund, Svensen et al (2001) studied fluid inclusions in veins (calcite-, quartz- and epidote-dominated types), and also investigated authigenic minerals grown in the Devonian sediments, away from the veins. The data allowed the authors to draw conclusions on the temperature and depth that existed during the formation of the veins: Solund: T = **305–330 °C**, burrial = **13.4 +/- 0.6 km**; Hornelen and Kvamshesten: T = **250 +/- 20 °C**, burrial: **9.1 +/- 1.6 km**.

Larsen (2002a, 2002b) studied quartz- calcite- and epidote-containing veins (as well as faulting and folding) in the northern part of the Hornelen basin, and interpreted these features to be a result of dextral transpression during westward extrusion (movement) of the basin along the original basin-controlling E-W trending marginal fault.

5.2.4 SUMMARY

Two major sedimentary facies are present in the HDM: (1) coarse conglomerates and breccias, and (2): sandstones, partly with intercalated conglomerate layers.

(1): More than **99 %** (by area) of the HDM consists of relatively coarse and monotonous alluvial fan/debris flow conglomerates and breccias characterized by almost complete absence of sedimentary structures. On the basis of difference in clast types, the conglomerates have been divided into two formations: the Vikafjellet Formation and the Blåfjellet Formation. The two formations are lateral equivalents and interfinger in the western parts of the HDM. In contrast, a peculiar breccia, the Litledokka Breccia, is probably a regolith or scree deposit.

(2): The fluvial sandstones constitute less than **1 %** of the HDM within the study area. Six small areas of alluvial fan/braided stream sandstones are present in the western- and northwestern-most areas. The sandstones are called the Mannen, "Galmannsskåra", Stegabruna, Novene, Vikaholmen and Gravaneset sandstone units.

Within the conglomerates, sandstone lenses also occur, generally between, but occasionally within, the debris flow beds. The bedding within the lenses, the lense shape and the debris flow margins are shown to be sub-parallel, i.e. representative of the overall subhorizontal bedding at the time of deposition.

Limited obsevations of palaeocurrent directions in distal sandstones located in the WSW, W and WNW/NW margin areas indicate that the transport directions may have been towards W or WNW in this area at this stage. At the SE margin, adjacent to the Nordfjord–Sogn Detachment Zone, the transport was possibly towards N.

The clasts in the HDM appear to consist of greenschist-facies rocks, and the source rocks for the clasts in the HDM thus appear to have occupied shallow-crustal positions during the Caledonian orogeny. Analogs to the ***types*** of rocks that were present in the original source area must therefore be sought in the Upper

Plate rocks to the west of the Nordfjord–Sogn Detachment Zone, although the actual source *areas* were probably located to the east of the present HDM, i.e. geographically within, but tectonostratigraphically above, the areas which are presently occupied by the eclogitic rocks of the Lower Plate. The essentially meta-sedimentary (meta-psammitic) clasts of the Vikafjell Formation may thus have been derived from rocks analogous to the rocks present in the substrate to the HDM. The essentially magmatic clasts of the Blåfjell Formation may have originated either from a Caledonian or a Precambrian plutonic complex.

The HDM are locally intruded by epidote-quartz dykes and quartz veins.

5.3 DEVONIAN/SUBSTRATE CONTACT RELATIONSHIPS

5.3.1 GENERAL

The present sub-chapter (Sect. 5.3) deals with the contact relationships along the base (margin) of the HDM. Two types of contacts will be discussed; *completely preserved* (100%) depositional angular unconformities, and *tectonically modified* depositional angular erosional unconformities. Emphasis will be on the first type, which is of interest due to its ability to display age relationships to structural elements in the substrate (Høydalsfjorden Complex, HC). Descriptions of depositional contacts along the intra-Devonian substrate *inliers* are included, although an overall description and structural interpretation of the inliers are postponed to Sect. 5.4.

The primary unconformities are important from three points of view:

(1) They reveal which deformation structures in the substrate that are cut by the sub-Devonian erosional surface, i.e. which structures that were formed prior to deposition of Devonian sediments. The present section (5.3) describes in more detail which substrate structures that are cut by the unconformity.
(2) They put important and rigid constraints on the analysis and interpretation of the three dimensional architecture of the HDM, and thereby also on the “original” basin configuration during times of deposition. This is discussed later (Sect. 5.5.8), but the present section contains the main documentation.
(3) They can potentially reveal the nature of the Devonian structures that have probably been superimposed on the substrate during the Devonian deformation (folding). However, it has not been possible to trace Devonian structural features across the unconformity, and this aspect is therefore not further treated in the present section.

The present secion starts with a presentation of the *completely preserved* primary unconformable contacts, including: overview and general contact features, a complete list of all localities with UTM-coordinates as well as a description of representative examples of contacts, and a description of palaeo-regoliths (Sect. 5.3.2). This is followed by a brief description of *tectonically modified* primary depositional contacts (Sect. 5.3.3). After a brief summary (Sect. 5.3.4), the significance of the primary unconformable contacts is discussed (Sect. 5.3.5).

5.3.2 PRIMARY UNCONFORMABLE CONTACTS

5.3.2.1 OVERVIEW AND GENERAL CONTACT FEATURES

The contacts around the margin of the HDM, as well as around the intra-Devonian substrate inliers, are accessable and have been followed by the author all the way along, unless otherwise indicated on the main map (**Plate 1**).

Table 5.2 contains a complete list of localities where the Devonian/substrate contact is clearly a primary depositional unconformity in the study area. The contacts are listed clockwise around the Devonian massif, starting with the areas in the west-northwest. The contacts separating the Devonian rocks and the intra-Devonian substrate inliers are included in the list. Along the Osen inlier, the localities are listed from west to east. The sites where primary unconformities are exposed are indicated by "contact-arrows" on the main map (**Plate 1**) as well as on the map sheets of **Appendix A**. On both of these, each "contact-arrow" generally indicates one exposure of a primary unconformable contact. However, at two map areas with a particular high concentration of localities, space problems on the maps implies that the number of "contact-arrows" is lower than the number of individual exposures in the field. This is the case at (1) Fjellsenden (localities no. 30-35, **Table 5.2**), where only 4 of 6 exposures are indicated on **Appendix A**. All localities at Fjellsenden are, however, indicated on **Plate 1.** Space problems on the maps is also present at (2) the Gravaneset area (localities no. 39-49, **Table 5.2**), where **Plate 1** contains only 5 of 11 exposures, and **Appendix A** only 1 of 11 exposures. All individual contacts at Gravaneset are instead shown on **Plate 4** and **Plate 5** where the full length of the primary contact-exposures is indicated by the contact line itself. On **Plate 1**, most of the field-walked (= “followed” in the map legend) Devonian/substrate contact is denoted "tectonically modified primary (angular) unconformity". Minor exceptions from this are found at Gravaneset, east of Straumsnes, and at the southern margin of the Osen inlier, where the contact is denoted "primary angular unconformity", since long distances along these contacts are without tectonic modifications.

The sites indicated as primary angular unconformities show practically no deformation post-dating Devonian ORS deposition. The unconformity-sites are separated by areas in which the contact is not directly exposed or in which the contacts exposed have been somewhat modified tectonically, yet still is more or less primary. The Devonian/substrate contact may, however, also be defined by *regular faults*; this is discussed below during the treatment of "tectonically modified primary unconformities" (Sect. 5.3.3) and later during the treatment of the faults in the HDM (Sect. 5.5.7.3).

The longest continuous exposures of the primary contacts are present at the Gravaneset area (**Plate 4** and **Plate 5**); along the southern margin of the Osen meta-psammite inlier; and at Fjellsenden. In these areas the exposures may be **5–10 m** long. In the other places, the unconformities are exposed over distances up to **a couple of metres**. The contacts may be present at the base of steep cliffs formed by the Devonian rocks (e.g. west and northwest of Mannen, north of Vikafjell, southeast of Nonsnova, south of lake Vassetvatnet, northern part of the Gravaneset area), or the surface of both the substrate and the Devonian rocks may be more equally and more shallowly inclined (e.g. top of Høgdene meta-psammite inlier, east of Vikanipa, south of the Teigafjellet Substrate Wedge, Fjellsenden, Kvangagjelet syenite, parts of the Gravaneset area). Neptunian dykes have not been observed with certainty, although possible neptunian dykes are present at localities in the Straumsnes area. Locally, the “flint”-hard yellowgreen epidote-quartz veins can be seen to cut the unconformity, but white quartz-veins cutting the unconformity have not been observed. Apart from the parasitic folds at Gravaneset (Sect. 5.5.3), none of the contacts are folded on outcrop scale.

5.3.2.2 EXAMPLES DOCUMENTING PRIMARY UNCONFORMABLE CONTACTS

The present section contains documentation of a selection of the primary unconformities listed in **Table 5.2**. The examples are taken from all around the HDM, and the description starts with the localities in the west-northwest and continues clockwise around the massif. The locality numbers given in each of the headings below correspond to the numbers in **Table 5.2**.

Contacts at the Straumsnes area (loc. no. 1-4; no. 3 documented)

Four exposures of an absolutely clear primary depositional unconformity are present in the Straumsnes area to the east of the road cut (**Plate 1**). Exposure no. 3 from the west will be documented below. All the contacts are situated along a **2–4 m** high cliff, and have a variable orientation due to an irregular palaeo-surface. For the three westernmost localities, however, the average contact is oriented with a ENE-WSW strike and a **50–70°** dip to the SSE (**Plate 1**). This is about the same dip as in the Gravaneset area (loc. no. 39–49, **Table 5.2**). Devonian pebbles are present adjacent to a substrate which mainly consists of meta-semipelitic rocks (**Fig. 5.34**). Devonian bedding in the Straumsnes area is oriented about **060/64 SE** and **071/60 SE (Plate 1**), and is thus generally parallel to the local contact surface, indicating that the contact was part of the original basin "floor". The basin has thus had a northwestward continuation. The substrate is generally made of intercalated layers of meta-greywacke and meta-semipelite, together with intrusive meta-gabbro (Høydalsfjorden Complex). F_2-folds with ESE-WNW trending fold axes are cut by the unconformity, as can be illustrated from other localities (see below). It may be noted that substrate S_1-foliations with an orientation of e.g. **078/30 S** are sharply truncated at the eastern-most exposure, although the S_1-cleavage can also have other orientations depending on the orientation of the F_2-fold limbs.

Contact to the northwest of Mannen (loc. no 6)

The **100 %** primary depositional contact is located in the northwestern corner of the HDM (**Plate 1**), and is presented in detail because it clearly demonstrates the structures in the substrate that are cut by the unconformity.

The contact is oriented with a strike/dip of **072/42 S**, and this measurement is based on the average palaeo-surface. The orientation of the contact is essentially parallel to the quite steeply dipping Devonian bedding in the area, indicating that this contact also originally formed a semihorizontal local basin "floor" prior to folding of the HDM. The substrate consists of strongly and continuously foliated (S_1) and folded (F_2) meta-semipelites and meta-greywackes (Høydalsfjorden Complex). The unconformity cuts an F_2 fold-hinge with orientation **123/20** and an axial plane oriented **050/32 SE** (**Fig. 5.35**). The F_2-folds are also truncated at a number of other places, notably in the Gravaneset and the Straumsnes areas. The Devonian material at this particular locality is coarse boulder conglomerates. One block at the locality has a size of **1.7 x 1.2 m** (**Fig. 5.35**). Elsewhere in the area, alternating layers of sandstone and pebbly conglomerate are present. The clasts consist of more than **90 %** whitish foliated arkosic meta-psammites.

Contact on top of the Høgdene meta-psammite inlier (loc. no. 8)

The depositional contact is located between the Høgdene meta-psammite inlier and the overlying "Galmannsskåra" sandstone unit **(Plate 1**). The primary contact at the "Galmannsskåra" sandstone unit was briefly mentioned in connection with the description of the sandstone unit (Sect. 5.2.2.2). The contact is **100 %** primary/depositional. Breccia clasts of foliated meta-psammite are present on top of the Høgdene meta-psammite and the clasts are identical to the foliated Høgdene meta-psammite. Reddish sandstone of the "Galmannsskåra" sandstone unit is present between the clasts (**Fig. 5.20**). The S_1-foliation, which is folded by the F_2-folds near by, is cut by the unconformity (**Fig. 5.20**). The orientation of bedding in the Devonian rocks is **056/57 SE**. The strike of the contact is similar to the strike of the bedding, but the dip of the contact is very difficult to discern; although it probably dips shallower than the bedding towards SE.

Contact to the east of Vikanipa (loc. no. 11)

The primary depositional unconformity to the east of the summit of Vikanipa (**Plate 1**) cuts meta-psammitic rocks in the substrate. An open crack up to **10 cm** wide locally separates the Devonian rocks from the substrate (**Fig. 5.36**), but sign of movements have not been observed. Devonian clasts are, however, observed in a **100 %** primary depositional contact against the substrate in a "pocket" on the eastern side of the crack. The easternmost, and N-S directed, Devonian/substrate contact appears to be quite steep as judged from the open gap and local minor displacement-zone fabrics further north. The Devonian bedding is oriented 091/68 S, thus standing with a high angle towards the contact. "Back-rotation" of bedding to original sub-horizontal position would still make the Devonian/substrate contact a steep cliff, showing that the present contact is a rotated palaeo-cliff (see Sect. 5.5.8). In the N-dipping steep slopes just to the north of the locality, greyish ultramylonitic substrate rocks with a W-E striking and S-dipping mylonite fabric, and with a thickness of **several 10's of metres**, approach the contact (Sect. 4.3.7.1). The S_1-mylonites can be traced until **a few metres** from the Devonian conglomerates, but the presice contact is not exposed. It is, however, obvious that also these mylonites have been cut by the Devonian unconformity.

Contacts to the southeast of Nonsnova (loc. no. 12–13; no. 12 documented)

The Nonsnova area is included here because it is one of the few places where the depositional unconformity can be seen to clearly cut the S_1-ultramylonites in the substrate. Two separate exposures are present to the southeast of Nonsnova (**Plate 1**), and at both of them the Devonian conglomerates cut ultramylonitic substrate rocks. The northern-most one will be described more closely (**Fig. 5.37**).

The Devonian conglomerate is present as an eastward-"pointing" **1–2 m** long wedge going down into the substrate, and with clasts consisting essentially of meta-psammite. It should be noted that the wedge is oriented at "right angle" to the general contact at this locality which has a tectonically modified contact striking essentially N-S and with a dip around **50° W**. The orientation of Devonian bedding is **070/59 SE** just to the west of the locality (on top of the Nonsnova summit). In the wedge, the conglomerate clasts are typically around **10 cm** in diametre, but range up to **15 cm**. The adjacent substrate mylonite has a typical grey or greenish grey colour. The Devonian/substrate contact is clearly a primary depositional unconformity, as the S_1-mylonitic foliation is sharply truncated by the Devonian conglomerate. However, it must be mentioned that along the contact between the two rock types, there is indications of a vague contact-parallel cleavage in the outermost **10 cm** of the Devonian conglomerate. No deformation can be observed mesoscopically in the inner part of the wedge. The position of the vague cleavage along the wedge margins suggests that the cleavage is related to a small degree of competence-controlled shear during the Devonian deformation. The fabric is only minor, and the mylonite foliation is clearly cut by the contact. The substrate-mylonite which is cut by the unconformity is a very common rock in the substrate, as described in Sect. 4.3.7.1. The other one of the two localities in this area is **100 %** primary with no sign of minor shear-related cleavage along the contact.

Contacts at the Teigafjellet Substrate Wedge (TSW) (loc. no. 14–15; no. 14 documented)

Two exposures showing **100 %** primary depositional unconformable contact relationships are present along the southern margin of the Teigafjellet Substrate Wedge (TSW) (**Plate 1**); the one to the west will be documented here. The Devonian conglomerates at this locality contain clasts mainly of meta-psammite, although some magmatic clasts are also present. Such a clast population is compatible to the position just within the transtional zone between the Vikafjell and Blåfjell Formations (Sect. 5.2.2.1). The substrate rock is a strongly foliated (S_1) meta-semipelitic to meta-psammitic rock. The primary unconformity strikes ESE-WNW and dips steeply S (**Fig. 5.38**). Devonian bedding is oriented **071/56 SE**, thus standing with a small angle towards the contact.

Contacts along the margin of the Osen inlier (Loc. no. 16–28; no. 28 documented)

A total number of **13** exposures of **100 %** primary depositional unconformities are present along the margin of the Osen meta-psammite inlier (**Plate 1**); **11** exposures along the southern margin and **2** along the western margin. As an example, the eastern-most locality of the southern margin will be documented here. The conglomerates of the transitional zone between the Vikafjell and Blåfjell Formations cut F_2-chevron-folded, whitish meta-psammites (**Fig. 5.39**). The Devonian/substrate contact generally appears to dip steeply towards S (**Fig. 5.40**), and with a bedding orientation of **076/60 SE**, the contact is parallel to the bedding, suggesting that the contact represents part of an original basin "floor".

Contacts at Fjellsenden (loc. no. 30–36; no. 30 documented)

Six examples of **100 %** primary depositional unconformable contacts are present at Fjellsenden in the area around the small wedge of substrate rocks going westward "into" the Devonian rocks (**Plate 1**). Four exposures are present on the slope going northwards from this wedge towards the cliff going down to Svardal, and two exposures are present along the margin of the wedge itself. F_2-folds are clearly cut by the unconformity (**Plate 1**). The northernmost exposure in the slope towards Svardal will be documented here. The magmatic clasts of the Blåfjell Formation give the conglomerate a massive appearance compared to the foliated (S_1) meta-psammites in the substrate of the Fjellsenden area (**Fig. 5.41**). The Devonian bedding in the area is oriented **151/63 NE**. To the north of the wedge, the Devonian/substrate contact appears to dip towards the N, NW or W – a more certain orientation is difficult to obtain due to lack of three dimensional control. To the south of the wedge, the contact is possibly more sub-horizontal, i.e. forming part of the original basin "floor". In the northward slope towards the cliff going down to Svardal, **a few metres** of the unconformity are inverted, i.e. substrate rocks are lying on top of the Devonian conglomerates. This feature is interpreted as a small and local palaeo-overhang in an originally steep contact surface. The deformation of the originally sub-horizontal Devonian bedding has rotated the bedding to its present SSE-NNW strike and quite steep dip. A "back-rotation" of the bedding to a sub-horizontal position in the Fjellsenden area would still leave a palaeo-overhang and palaeo-cliff in the substrate.

Contact at the Kvangagjelet syenite (loc. no 37)

A **100 %** primary depositional unconformity is present in the river coming from the Håsteinsvatnan lakes (**Plate 1**). The contact appears to dip steeply towards N (**Fig. 5.42**), i.e. sub-parallel to the Devonian bedding which is oriented **154/59 NE**, implying that the contact was part of the basin "floor". Pre-Devonian cracks in the sub-Devonian palaeo-surface are present at the unconformity (**Fig. 5.43**). The clasts in the conglomerates of the Blåfjell Formation (see Sect. 5.2.2) are of the same type as the rocks of the Kvangagjelet syenite.

Contacts at the Gravaneset area (loc. no. 39–49; no. 41 and 48 documented)

A total number of **11** exposures of the sub-Devonian **100 %** primary depositional unconformity are present in the Gravaneset area (**Plate 4** and **Plate 5**), and localities no. 41 and 48 will be closer described as examples.

Locality no. 41 is located at the western margin of the large "pocket" of Devonian rocks that have been faulted into the substrate by the fault shown on **Plate 5**. (The locality can also be seen on the photo of **Fig. 5.27** at the uppermost of the contact arrows). At this locality, interlayered sandstones and gravelstones/conglomerates of the Gravaneset sandstone unit cut substrate rocks that on outcrop-scale are both folded (F_2) (**Fig. 5.44**) — and at a site **a few metres** further north — unfolded (**Fig. 5.45**). The Devonian bedding is oriented around **015/58 E**, which is sub-parallel to the local contact, and the contact thus represents a local original basin "floor".

Locality no. 48 is located in the road-cut at the southern end of the Osstrupen road-bridge on the northeastern side of the Gravaneset area (**Fig. 5.46a**). At this locality, a **20 m** high vertical road-cut exposes the eastward dipping unconformity. The contact is only slightly modified tectonically (see Sect. 5.3.3), but in the lower middle-part of the road-section, conglomerates resting with a **100 %** primary depositional unconformity against the substrate rocks form a "pocket" into the substrate (**Fig. 5.46b** and **5.46c**). The Devonian bedding is oriented about **030/61 ESE**, which is sub-parallel to the local contact. The contact is thus part of the local original basin "floor". On top of the road-cut, the conglomerate rests on meta-gabbro, and the unconformity is cut by a set of thin epidote-quartz-veins (**Fig. 5.47**).

5.3.2.3 REGOLITH DEVELOPMENT

A regolith is in the present work, taken to mean a sedimentary rock (e.g. a breccia) which lies essentially in situ or has been only slightly moved with repect to its original site of formation, having been formed essentially from the immediately underlying substrate. Regolithic rocks may be present intermittently on unconformable (palaeo-) surfaces and represent a stage in the development of a palaeosoil.

In the study area, regoliths are a rather uncommon feature. Where present, the regolith may have two modes of occurrence; it is either present at the very contact between Devonian rocks and substrate rocks (type 1), or it is present *within* substrate areas where the once overlying Devonian sediments are now eroded away (type 2), leaving patches of regolith behind as outliers. Examples of both these variants are presented below.

Type 1-regolith

Along the precise contacts of the HDM, regolithic breccias are rare. The only place where a possible regolith of this type occur, is at the boundary between the Høgdene meta-psammite inlier (Sect. 5.4) and the "Galmannsskåra" sandstone unit. At this locality, clasts of foliated meta-psammite at the contact are completely angular, and red sandstone of the unconformably overlying sandstone unit is situated between the clasts (**Fig. 5.20**). Some small amount of transport may have occurred, but the clasts directly reflect the immediately underlying meta-psammites of the Høgdene meta-psammite inlier.

Type 2-regolith

At Svardal, a breccia is present within an area of substrate rocks. It is located on the small penninsula going northwestwards into the lake Svardalsvatn, as seen from Svardal. The area is located **300 m** from the nearest "proper" Devonian rocks (**Plate 1**). The breccia at this particular locality covers an area of **a few square metres**, and is surrounded by undisturbed (unbrecciated) substrate rocks of foliated (S_1) and folded (F_2) semipsammitic schists **a few metres** away. The clasts are variably oriented (**Fig. 5.48**). The breccia fabric appears to change gradually into undisturbed foliated rocks towards the "margin" of the breccia. The breccia is clearly not fault related, since there is no fault zone running through the area. The position of the breccia, the type of clasts, and the angular clast shapes indicate that it is not an ordinary, transported Devonian deposit, but a regolith that is sitated in situ where it was derived. The "proper" Devonian conglomerates were probably situated just above this locality, but have later been removed by erosion.

The Litledokka Breccia on the plateau between Storedokka and Litledokka (**Plate 1**) was described in Sect. 5.2.2.1. The Litledokka area apparently consists of a breccia with clasts of a type which directly reflects the local meta-sedimentary substrate. The clasts are very angular, without any sign of rounding (**Fig. 5.17**). The contact between the breccia and the undisturbed substrate rocks has not been observed. The clasts in the breccia are completely randomely oriented. Although parts of the rock may have experienced some degree of transport, large parts appear to have a regolithic character. A contact between the breccia and "proper" Devonian deposits are exposed **400 m** to the east-southeast of the summit of Storedokka. The contact is sharply defined (**Fig. 5.18**), and the Devonian deposits are characterised by variably ***rounded*** pebbles and the occurrence of magmatic clasts (syenites, etc.) that are "exotic" to the immediately underlying substrate. The "proper" Devonian deposits thus have an appearance which is distinctly different from the regolithic breccia.

5.3.3 TECTONICALLY MODIFIED PRIMARY UNCONFORMABLE CONTACTS

General

Tectonically modified versions of the primary depositional unconformable Devonian/substrate contact appear to be quite common in the study area, and brief field descriptions of a few examples will be presented in the following.

Along the Devonian/substrate contact, movement zones may generally be of a wide range of types from semiductile to brittle, from thin (**0–5 cm**) to thick (**30–40 cm**), and with modest to more intens fault-fabric and fault rock development. In addition to the movement zones which are here classified as "tectonic modification zones", and which produce the "tectonically modified primary unconformable contacts", another category movement zones, classified as *regular faults*, are apparently also present along the contact, and the following distinction is therefore made between "regular fault" contacts and "tectonically modified primary unconformable" contacts: The zones which are more strongly developed, and which tend to have steep to vertical dips and thus in most cases appear to be *oriented independently of the usually shallower dipping Devonian/substrate contact*, are considered as regular faults, and the treatment of these faults is postponed to Sect. 5.5.7. The zones which are more modestly developed, and which also appear to follow and/or to be *oriented essentially parallel to the assumed or observed orientation of the Devonian/substrate contact (envelope) surface*, are considered to be "tectonic modifications" of the original primary unconformable contact, and are treated below. However, due to a general lack of control on the three dimensional orientation of the sub-Devonian contact, it may sometimes be difficult to discern whether the movement zone actually follows the contact at depth, and the distinction between the two displacement types are then difficult to make. The term "tectonically modified primary unconformity" is used in a nongenetic sense, i.e. without relating the modifications to specific larger-scale tectonic events, as it is difficult to relate the various minor movements in space and time.

Contact at Straumsnes

In the road-cut at Straumsnes (**Fig. 5.49a**) (UTM 0655 3063) (**Plate 1**), a *tectonically modified* primary Devonian/substrate contact can be observed (**Fig. 5.49b** and **5.49c**). The nearest one of the four *proper* primary depositional unconformities exposed in this area, are present only **35 m** to the east of this locality (Sect. 5.3.2.2). At the road-cut contact, where the Devonian conglomerate clasts are situated on top, the clasts are essentially undeformed all the way down to the precise contact, under which a **30 cm** wide zone of the substrate rocks (intercalated meta-greywacke and meta-semipelite layers) have been slightly tectonised parallel to the contact (**Fig. 5.49c**). The displacements may be recognised from the essentially contact-parallel, although anastomosing, shear fractures and glide planes which isolate poorly defined "fragments" and lenses in the zone, and form a coherent/cohesive, but very modestly developed, fault rock. According to the classification diagram of Braathen et al. (2004), the fault rock appears to range from cataclasite (produced from the greywacke component), to a more phyllonitic type (produced from the meta-semipelite layers). The movements took place under semiductile to semibrittle conditions. The orientation of the tectonic zone is **068/70 SE**, which is parallel to the Devonian bedding with orientation **060/64 SE** near by (**Plate 1**).

Contact south of the Osstrupen road-bridge

Another tectonically modified primary unconformity is present in the road-cut at the northeastern part of the Gravaneset area, just to the south of the Osstrupen road-bridge (UTM 0612 3047) (**Plate 1**, **Plate 4** and **Fig. 5.46**). The primary depositional unconformity at this locality was described in Sect. 5.3.2.2. The tectonised zone follows quite precisely along – and partly affects – the primary contact from the bottom to the top of the **20 m** high road-cut. The only major exception from this zone/contact trace-parallelism is found at the "pocket" of conglomerates going "down" into the substrate in a "palaeo-depression" (**Fig. 5.46b** and **5.46c**), where the tectonised zone cuts straight through the "neck" of the "pocket". Where the zone follows – and partly affects – the precise contact, the zone tends to be developed mainly within the basal part of the Devonian conglomerates. From the top of the road-cut, the zone appears to continue along the contact also in the lateral direction towards SW. The zone is generally **10–15 cm** thick and defined by a set of semibrittle to brittle shear fractures anastomosing along the contact and defining intrazonal "lenses" which may be up to **20 cm** long and **5 cm** thick (**Fig. 5.46c**). The orientation of the tectonised zone is **056/64 SE** (**Plate 4**), and the orientation of Devonian bedding near by is **030/61 SE**.

Contact at southern part of the Gravaneset area
(NB: tectonisation dated by palaeo-magnetism)

At the Devonian/substrate contact in the eastern road-cut at the southern part of the Gravaneset area (UTM 0578 3029) (**Plate 5**), the once primary unconformable contact has been subjected to minor tectonic modifications. This minor fault zone has a thickness of about **20 cm**, and contains a modestly developed cohesive fault rock, which has been preliminary classified as a protocataclasite. Torsvik et al. (1987) sampled the fault rock for palaeomagnetic dating of movements, and reported that the samples of the zone were taken where the zone is positioned just on the substrate side of the contact. The fault rock itself was not described. (Visual identification of the presice Devonian/substrate contact was difficult in this road section). The remanence bearing minerals were reported to be of essentionally two types: haematite at the *margins* of the zone, and secondary magnetite at the *central* parts of the zone, replacing haematite. Based on the palaeomagnetic pole positions, a poorly confined Triassic/Early Jurassic age was obtained for the fault movement. From dating of other fault zones in Western Norway, the authors concluded that although the precise ages were not well calibrated, it was evident that extensional faulting, notably reactivation of "Caledonian", and in particular "Devonian", structures in western Norway, was widespread during the Mesozoic. The orientation of the tectonised contact is **062/62 SE** (**Plate 5**), and the bedding near by is oriented **048/66 SE** (**Plate 5**).

Amount and time of displacement

The tectonic modification-"zones" (contact-parallel deformation zones) described above have not been observed to cut marker horizons. Although the amount of displacements on these "zones" are therefore uncertain, the very modest development of deformation fabrics and fault rocks indicates that the movements have been only very minor. When taking the large number of **49** unconformity-exposures (Sect. 5.3.2) around the HDM into consideration, the presence of only very local and minor movements implies that in the overall picture, the HDM as a whole is situated with a primary depositional unconformity on top of the substrate.

Generally, movements appear to have occurred under semiductile, semibrittle and brittle conditions, indicating that movements took place over a fairly large time span. Apart from the dated movement zone at Gravaneset (see above), it is very difficult to precisely estimate the time at which the various individual zones were active. Nevertheless, some general considerations can be made. The clearly brittle zones definitely post-date the Devonian D_1-deformation which produced folding and cleavage of the HDM. As to the semiductile to semibrittle zones, it is not known whether the zones originated during the Devonian D_1-deformation, but the possibility cannot be excluded, since the D_1-deformation (folding) may have produced slip due to competence contrasts at the Devonian/substrate contact. It must, however, be expected that later reactivations may have overprinted, to a varying degree, any such possible original Devonian D_1-related fabrics, suggesting that the fabrics observed now are partly later features, possibly Mesozoic in age.

5.3.4 SUMMARY

As much as **49 individual exposures** of the primary depositional unconformable Devonian/substrate contact have been located around the Håsteinen Devonian Massif. At all these contact localities, the main S_1-foliation in the substrate (Høydalsfjorden Complex) is cut by the sub-Devonian unconformity. F_2-folds are also clearly cut by the unconformity at the localities Gravaneset, Straumsnes, Mannen, along the southern margin of the Osen meta-psammite inlier and at Fjellsenden. S_3-ultramylonites are truncated at the localities to the east of Vikanipa and also to the southeast of Nonsnova. The primary unconformity is sub-parallel to the Devonian bedding at several localities, for example at Straumsnes, northwest of Mannen, southern margin of the Osen inlier, Fjellsenden, Kvangagjelet syenite and Gravaneset, implying that the contact was originally part of the local basin "floor". Regoliths are rare, but two types occur: either the regolith lies directly at the Devonian/substrate contact with proper Devonian sediments on top; or the regolith occur within an area containing only substrate rocks, where the once overlying Devonian rocks have been eroded away.

Tectonically modified primary Devonian/substrate contacts are common and may, for example, be observed in the following three road-cuts; at Straumsnes; to the south of the Osstrupen road-bridge; and at the southern part of the Gravaneset area. Very modest development of fault rocks shows that the movements are only local and minor, and does not alter the fact that the HDM, in the overall picture, is sitated *in situ* with a primary unconformable contact on the substrate of the Høydalsfjorden Complex. The movements probably occurred over a large time span. Semibrittle to brittle fabrics post-date the Devonian D_1-deformation. When it comes to the semiductile to semibrittle fabrics, it is unknown whether they originated during the Devonian D_1-deformation. The fabrics *now* observed, however, may partly be a result of later overprinting displacements, since reactivations must be assumed to have occurred. Such movements may have been particularly common during the Mesozoic.

5.3.5 DISCUSSION: SIGNIFICANCE OF THE UNCONFORMITIES

Relative age of substrate structures

The fact that the unconformity truncates the typical HC S_1-foliation and F_2-folds has important consequences for the deformational history of the area. The D_1-deformation that produced the main foliation, and the D_2-deformation that produced the folds have been interpreted as related to the Scandian top-to-the-east contractional phase of the Caledonian orogeny. The dominating fold structures in the Høydalsfjorden Complex are thus older than the deposition of the Devonian sediments, i.e. pre-Middle Devonian.

Although "contractional " top-E mylonites have not been observed, it is, theoretically possible that "contractional " mylonites are present in the substrate, and that these mylonites have been overprinted by the "extensional" top-W mylonite (see Sect. 4.3.4.1).

The unconformity also cuts the HC S_3-mylonites, implying that a *pre-Middle Devonian mylonite generation* is present in the substrate. The mylonites are top-to-west "extensional-mylonites" formed during the Devonian (post-Scandian) crustal-scale extensional movements ("orogen-collapse") which produced (and deformed?) the Devonian basins (e.g. Seranne et al. 1989; Fossen 1992; Osmundsen & Andersen 1994). The S_3-mylonites are possibly related to the Mode-I extension, as defined by Fossen & Dunlap (1998).

In-situ position of the HDM

Another important consequence of the presence of the completely primary depositional unconformities all around the HDM is that they show that the HDM is situated *in situ* with respect to its original basin "floor". This means that the HDM has not been thrust ("Scandian contraction model") or detached ("Devonian detachment model") with respect to the ***immediate*** substrate of the basin.

The "detachment model" states that the **several km** thick mylonitic Nordfjord–Sogn Detachment Zone (NSDZ) is present somewhere below the HDM/HC package (as it is below the other Devonian massifs in Western Norway). Regionally, the NSDZ is believed to have had a gentle westward inclination, and it can be of interest to consider how high up towards the base of the Håsteinen Devonian deposits the mylonite zone is developed. The depth to the subjacent mylonites of the NSDZ naturally depends on the amount of dip-slip on the two bordering brittle faults, the Sunnar Fault on the north side and the Standal Fault to the south. If the HDM/HC block has experienced a relative drop of e.g. **1 km** compared to the NSDZ-mylonites, then the detachment is found well below the present HC-surface, and the exact depth to the HC-surface would depend on the unknown thickness of the HC. The investigations show that the HC rocks in the western part of the study area have not been effected by NSDZ mylonitisation, and the zone must, hence, go well below this area. If the NSDZ was generally rising eastwards, the Svardal area, and the area along the lake Vassetvatnet, would be the part of the study area that would be located closest to the "detachment zone" However, as documented by Wilks & Cuthbert 1994, the mylonite fabric in the NSDZ between Håsteinen and Hornelen has a sub-horisontal envelope surface, suggesting that the NSDZ is not positioned shallower, i.e. closer to the HC rocks, in the eastern part of the study area. The presence of a regolith in the Svardal area, and the general lack of mylonites of the general "detachment"-type, suggest that in neither of these areas has the "detachment"-related shear been developed strongly and shallowly enough to affect the basin "floor" below the HDM. This means that the shear fabrics related to the "Nordfjord-Sogn Detachment Zone" (present in the Eikefjord–Sunnarvik–Osa area of the present thesis) must lie at a deeper level below the HDM/HC area. Since the depth to this zone is not known, the zone has not been drawn into the profiles across the study area (**Plate 2**).

A comparison of the sub-Devonian contacts of Håsteinen with the ones of Hornelen is useful. The fact that the steeply dipping beds of the HDM lies unconformably on top of Upper Plate rocks, instead of being in direct tectonic contact with the detachment zone as in the Hornelen Devonian Massif (e.g. Seranne et al. 1989), has serious consequences for the structural modelling of the HDM. The Hornelen sediments were deposited aginst a basin-controlling listric brittle fault in the east, a fault which might have been the continuation of the NSDZ appearing at the subaerial surface. As the basin floor moved westwards, the beds experienced roll-over rotation and eventually came into contact with the mylonite level of the zone. However, it should be noted that the *present* – eastern contact of the Hornelen deposits is presumably a Mesozoic fault, that has juxtaposed the Devonian sediments against structurally even lower parts of the NSDZ mylonites. These differences between Hornelen and Håsteinen has significant implications, that is discussed in Sect. 5.5 and Ch. 6.

Original size of the basin

The presence of primary unconformities along the present southwestern, western and northern margins of the HDM indicates that the basin originally had a wider extent on the large areas of HC rocks in these directions. (The eastward continuation of the massif is represented by the deposits presently situated to the east of the lake Vassetvatn, deposits which were not included in this study). Hence, the original size of the Håsteinen basin was significantly larger than the present size. Along the southeastern margin, however, the situation is different. Here, the HDM is located adjacent to the Nordfjord–Sogn Detachment Zone (Standalen segment, Andersen & Jamtveit 1990), with its northward-dipping mylonite foliation. The brittle Standalen Fault, which possibly formed in the Mesozoic, follows along, and cuts, the NSDZ mylonites. As discussed in Sect. 5.2.2.3, the ***present-day*** southern margin of the HDM is probably also located close to the ***original*** margin of the Devonian basin. The Standalen Fault is probably situated close to the position of the original *Devonian* brittle fault that controlled the deposition of the Håsteinen sediments.

The northern margin of the Håsteinen deposits faces the Hornelen sediments, and the two are separated by a distance of **10 km**. The primary unconformity along the northern margin of Håsteinen shows that the basin had a larger size in the northward direction. In Devonian times, however, it is not likely that the HDM coalesced with the Hornelen Devonian Massif across the present "gap", since the Hornelen basin was fed from Upper Plate source rocks positioned to the south i.e. positioned between Hornelen and Håsteinen basins. However, it cannot be excluded that the basins coalesced further west, possibly at a higher stratigraphic levels. Thses issues are further discussed further in Sect. 6.4.3.

5.4 INLIERS OF SUBSTRATE ROCKS WITHIN THE HDM

5.4.1 GENERAL

Sect. 5.4 deals with the inliers of substrate rocks present within the HDM. The term "inlier" is here used in a non-genetical sense in accordance with Bates & Jackson (1980) as denoting "an area or group of rocks surrounded by rocks of younger age". The term "outcrop" is taken to mean the *whole area* where a rock type could potentially be exposed if there were no Quaternary cover. The term "exposure" is taken to mean the *particular places* where the rocks are exposed subaerially.

The HDM contains the following inliers: the Høgdene meta-psammite inlier (Sect. 5.4.2), the Osen meta-psammite inlier (Sect. 5.4.3), and the Litleteigen greenschist inlier (Sect. 5.4.4). In addition, the Kvangagjelet syenite (Sect. 5.4.5) will be discussed, since theoretically it may be located within the HDM. A brief summary is given at the end (Sect. 5.4.6). The general geology of the inliers was described in Sect. 4.2.1.4, and the primary depositional unconformities in Sect. 5.3.

The purpose of the present treatment is to interpret the occurrence of each of these inliers. The following models are relevant when such inliers are to be explained: they may (1) have a sedimentary origin, i.e. be landslides, (2) be due to the present sub-Devonian topography, i.e. be topographic highs, (3) have a tectonic origin, i.e. be cores of anticlines, or be highs due to imbricated slices or high angle reverse faults, etc.

The alternative that the inliers be synsedimentary horsts is considered very unlikely, since no signs of synsedimentary faulting has been observed in Håsteinen. The horst alternative is thus not further discussed in the section.

For each inlier, the geological features of significance for the interpretation is first described, whereafter the inlier itself is interpreted.

The interpretation of each inlier is partly based directly on data supporting the particular model favoured ("positive evidence"), and this interpretation and related data are presented first. But, the interpretation is also partly based on the rejection of alternative models ("negative evidence"), and these alternative models and the bases for rejecting them are discussed individually, subsequent to the interpretation of each inlier.

5.4.2 THE HØGDENE META-PSAMMITE INLIER

5.4.2.1 DESCRIPTION

The inlier

The Høgdene meta-psammite inlier is located along the northwestern margin of the HDM (**Plate 1**), and the outcrop is transected by the western profile (**Plate 2**). The name Høgdene is taken from the topographical 1:50.000 map "Eikefjord" (1118 II). The meta-psammite is located on a very steep northward dipping mountain slope, and the steep slope restricts accessibility. The W-E length of the inlier is about **1.3 km**, and the thickness (as measured on the hillside) is maximum **90 m** in the west and thins to **40–50 m** in the east. The width in the map plane is up to **100 m** in the west, and narrows towards the east (**Plate 1**). The inlier essentially consists of foliated meta-psammite of arkosic to quartzitic composition, but in the western areas, more pelitic rocks, as well as marbles, are present (Sect. 4.2.1.4). The meta-psammite has an appearance which is very similar to the substrate rocks outside the HDM, although the amount of whitish meta-psammite is larger in the inlier than in the surrounding substrate. With only limited parts of the inlier accessable, the structural development in the inlier has not been studied in detail. However, from local reconnaissance, it appears that the main S_1-foliation as well as the F_2-fold axes have essentially the same orientation as in the ordinary substrate outside the HDM. The meta-psammite is locally brecciated (**Fig. 4.19**). This breccia is not fault-related, as no fault zones are connected to the breccia. The breccia may cover areas up to **a couple of metres** wide, and has been observed to show a gradual transition into a continuously foliated rock. All clasts in the breccia are derived from the surrounding foliated meta-psammite; no exotic clasts are present in the breccia, which is interpreted as a Devonian regolith.

The contacts

Since the inlier crops out in a steep hillside and has an elongated shape laterally along the hillside, the inlier can be said to have a "lower" and an "upper" contact in this hillside. The "lower" contact, which is towards the Devonian Stegabruna sandstone unit, is exposed at two localities. These are located at the lake at Høgdene, and at "Galmannsskåra". At both these localities the contact is interpreted as a primary depositional unconformity. At "Galmannsskåra", the contact is **100 %** primary, but at the Høgdene lake a very small movement might have occurred, as a **1 mm** thick possible "fault rock" is present. Any movement must, however, be so minute that it may surely be ignored in the present context, and the contact is considered as a primary unconformity. The "upper" contact of the meta-psammite, which is towards the sandstones of the "Galmannsskåra" unit, is exposed at one locality situated **50 m** to the east of "Galmannsskåra". Also this locality is **100 %** primary without any sign of tectonic modification. (An additional locality is also possibly locateted at the very eastern end of the inlier, but this is uncertain and the locality is therefore not listed in **Table 5.2**). Both the lower and upper contacts of the Høgdene meta-psammite inlier are thus interpreted to be primary deposional unconformities.

Adjacent Devonian sediments

The Devonian sediments located adjacent to the inlier do not show signs of increased tectonic disturbance. This indicates that the areas around the inlier have not undergone more deformation than other areas in the HDM, or experienced any particular kind of deformation related to the nearness to the inlier.

The contact-bedding relationship

The bedding in the Devonian rocks around the inlier (i.e. in the Stegabruna sandstone unit, the "Galmannsskåra" sandstone unit and the conglomeratic Vikafjell Formation) everywhere dips moderately to steeply towards the SSE, and have a strike oriented WSW-ENE. The bedding orientation "below" and "above" the inlier (in the hillside) is therefore essentially parallel. The orientation of the original Devonian/inlier contact that once continued across the present inlier outcrop has been reconstructed from the outcrop pattern and the presence of breccia (interpreted as regolith). This contact must have had a general but irregular dip towards the north (see **Plate 2**, western profile), with a high angle to the SE-ward dipping bedding.

5.4.2.2 INTERPRETATION

Based on the features described above, the Høgdene meta-psammite inlier is interpreted to be an outcrop of "rooted" substrate forming a sub-Devonian topographical high. This means that the erosion has revealed the substrate below the cover of Devonian sediments. The long and thin shape of the inlier is interpreted to be determined by the original shape of the high and the present topography (erosion level). The breccia on the surface of the meta-psammite is interpreted to be a regolith, indicating that the cover was present just "above" the present N-ward-dipping erosion surface. The presence of primary depositional unconformities on the hillside both "below" and "above" the inlier is consistent with the above interpretation of the inlier.

Since the interpretation above is also partly based on the rejection of alternative models, it is important to explain in some detail why these models cannot be applied to the Høgdene inlier, and they are therfore discussed individually below.

(1): Landslide: Landslides have previously been reported from two of the Devonian massifs of western Norway; the Solund Massif (Michelsen 1986; Michelsen et al. 1986; Michelsen et al. 1983a; Michelsen et al.1983b; Hartz et al. 2002), and the Kvamshesten Massif (Kolderup 1923; Skjerlie 1971; Markussen & Svenby 1992; Osmundsen 1996; Osmundsen et al. 1998). As seen from the maps (**Plate 1** and **Appendix A: maps no. 6a** and **6b**), the WSW-ENE strike direction of the Devonian bedding is sub-parallel to the "map trend" of the "lower" contact of the Høgdene inlier. Despite this sub-parallelism, the bedding strikes slightly more northeast-wards than does the trend of the inlier-contact on the map, which appears to indicate that they are not really parallel after all. However, if the "map-trend" of bedding could have been traced on the N-dipping hillside, the intersection of the bedding and the topography would have given a bedding-"map-trend" even ***more*** parallel with the contact-"map trend" than what appears from the map, and the bedding and contact must thus be considered as parallel. Since the base of a landslide would tend to be parallel to the subjacent bedding, this situation could comply with a landslide model. Also the ***dips*** of the contact and bedding should be fairly parallel to each other to make the landslide model likely, but the dip orientations are more uncertain in the present case. However, the presence of primary contacts both "above" and "below" the inlier makes landsliding unlikely. Furthermore, the brecciation in the inlier is probably not related to "shaking" during sliding, since the breccia is surrounded by undisturbed rock, and "shaking" is thought to produce zones of interconnected brecciation. Along the base of the meta-psammite outcrop there is no signs of slide-related features like sand or conglomerate "dykes" intruding into the meta-psammite, or a higher degree of "shake"-brecciation, etc., which have been reported as an important diagnostic criteria for the Hersvik landslide in the Solund Devonian Massif (Michelsen et al. 1983b; 1986; Hartz et al. 2002). The amount of brecciation throughout the rest of the inlier is low, also suggesting that the inlier is not a landslide. In addition, the Høgdene inlier is very thin compared to its length, and would probably have broken into pieces if it was a "landslide", but such fragmentation has not been observed. Although the present inlier-outcrop is long and thin, it is theoretically possible that larger parts of a possible "landslide" might be covered by Devonian sediments, but indications of this have not been found and the idea is difficult to test. In conclusion, it appears that the landslide model cannot be applied to the inlier.

(2): Tight anticline: The possibility that the outcrop is the core of a tight anticline that has been revealed due to erosion can be rejected. The strike and dip of bedding in the deposits on the top and at the base of the outcrop are essentially the same. If the outcrop should be the core of an anticline, this would imply an isoclinal fold in the Devonian, i.e. with completely transposed limbs. The Devonian sediments around the inlier display no signs of axial planar cleavage and/or parallel orientation of pebbles due to flattening/reorientation, or any other deformation. It is highly unlikely that isoclinal folding could take place without producing any clear signs of deformation in the Devonian rocks, hense the anticline model is rejected.

(3): Imbrication (low angle/high angle reverse faulting) of "basement-cover": The hypothesis that the inlier is related to imbrication of a substrate-cover unit along a tectonic lower contact must be rejected since the lower contact is a primary unconformity and not a tectonic "thrust" zone.

5.4.3 THE OSEN META-PSAMMITE INLIER

5.4.3.1 DESCRIPTION

The inlier

The Osen meta-psammite inlier is situated in the central eastern part of the HDM (**Plate 1**), and is transected by the central profile (**Plate 2**). The name is taken from the topographical 1:50.000 map "Eikefjord" (1118 II). At Osen, the meta-psammite is covered by large deposits of Quaternary deltaic marine terraces which divide the inlier into two parts. Most of the meta-psammite is present to the east of the Quaternary deposits, but a small area crops out on the western side. The length of the meta-psammite is **2.5 km**, and the width up to **0.6 km**. In the easternmost parts, the meta-psammite disappears below the Devonian cover to reappear as an eastern continuation in the Teigafjellet Substrate Wedge (TSW). The cover of Devonian rocks on top of the substrate rocks between the inlier and the TSW must be fairly thin (**~50 m** ?). Models to explain the presence of the Osen meta-psammite inlier must also explain the presence of the TSW.

The Osen inlier generally consists of a whitish foliated arkosic meta-psammite. At Osen, in the areas just to the northeast of the bridge crossing the river to the east of the Osen Farm, a very small area of more meta-pelitic rocks and marbles is present (Sect. 4.2.1.4). The Osen meta-psammite is locally brecciated. The breccia-areas have irregular shapes and may be up to **a couple of metres** wide. The clasts are completely angular (**Fig. 4.22**), and it has been observed that the clasts gradually change into continuously foliated rock at the margin of the breccia-areas. All clasts in the breccias are derived from the surrounding meta-psammite, no exotic clasts are present. The breccia occurs as isolated patches surrounded by continuously foliated meta-psammite, and this means that the breccia has not been formed by fault zones, but is to be interpreted as regolith.

The Osen inlier consists of monotonous foliated meta-psammite of the same type as is present in the local substrate outside the HDM. In the substrate close to the outer margin of the HDM, most areas of meta-psammite are of smaller size, but meta-psammites of similar size are present between Kalsvik and Sandvika to the northwest of the study area (see the geological map of Kildal 1970). The Osen meta-psammite is of the same type as that previously described in the Høgdene meta-psammite inlier (Sect. 5.4.2). As shown in **Fig. 4.45**, the orientation of the structural elements in the Osen inlier is very similar to the orientation outside the HDM; i.e. the S_1-foliation in the inlier strikes essentially W-E and dips steeply mainly to the S; the F_2-fold axes trend essentially W-E with shallow plunges mainly towards the east; and the axial planes to the F_2-folds strike W-E with steep dips both towards S and N.

Contacts

The southern contact of the inlier is exposed in at least 11 localities (**Plate 1**), and all of these are **100 %** primary depositional unconformities (Sect. 5.3.2) (**Table 5.2**). The western contact is exposed at 2 localities, both of which are also **100 %** primary (**Plate 1**, **Table 5.2**). The northern contact is exposed at several localities, as a fault. The fault is described in more detail in Sect. 5.5.7, where it is concluded that the displacement on the fault is only minor, i.e. in the range of **50 m**, with an essentially dip-slip (reverse) movement. The contact between inlier rocks and Devonian sediments goes partly on the southern side of the fault (**Plate 1**), showing that the juxtaposition of the inlier and the Devonian rocks is not controlled by the fault, and the fault is thus only to a small degree responsible for the presence of the inlier.

Adjacent Devonian sediments

The Devonian sediments adjacent to the inlier display no signs of tectonic disturbance, and the areas around the Osen meta-psammite have not been subjected to more deformation than the other areas in the HDM.

The contact - bedding relationship

The strike of bedding in the Devonian deposits comes into the northern margin of the meta-psammite with an angle of **20–30°**. Along the southern margin, the bedding-strike is margin-parallel in eastern parts, but make an angle in the southwestern part. The bedding shows a constant dip of **50–70°** to the SSE over the whole area and on all sides of the inlier, i.e. the irregular outline of the inlier is unrelated to the Devonian bedding structure, and the Devonian/inlier contact is generally discordant (i.e the sub-Devonian unconformity formed a palaeo-relief). The angle between bedding and the original Devonian/inlier contact that was once present across the inlier, must have been generally high, although variable (see central profile of **Plate 2**, and discussions in Sect. 5.5.8).

5.4.3.2 INTERPRETATION

Based on the data presented above, the Osen meta-psammite inlier is interpreted as an outcrop of "rooted" substrate forming a sub-Devonian topographic *high* that has been revealed through erosion. The breccias present on the surface of the meta-psammite is interpreted as local regoliths, indicating that the Devonian cover was present just above the present erosion surface. The fault along the northern margin of the meta-psammite has contributed to the exhumation of the meta-psammite by lifting it up relative to the area to the north of the fault, but the displacement has been only minor, and thus not a controlling factor for the presence of the inlier. The similarity in orientation of structural elements in the inlier and in the ordinary substrate outside the HDM, indicates that the inlier is "rooted" in the substrate. This is also confirmed by the fact that the Teigafjellet Substrate Wedge (TSW), which represents the eastern "continuation" of the inlier, is part of the ordinary substrate.

As for the Høgdene inlier, the interpretation above is is also partly based on the rejection of alternative models, and these are therefore discussed below.

(1): Landslide: The Teigafjellet Substrate Wedge (TSW), which is the eastern continuation of the Osen meta-psammite inlier, appears to be an integral part of the substrate. If the meta-psammite was a landslide, the "plate-like" and bedding-parallel TSW should, at its "base" in the hillside to the east, be separated from the substrate, possibly with Devonian sediments between the substrate and the TSW. Although exposures are limited, no indications of such a separation is present in the hillside to the east of the TSW, and the nature of the TSW therefore suggests that the Osen inlier cannot be interpreted as a landslide.

Furthermore, the general discordance between the regular Devonian bedding and the contact along the northern margin of the inlier also speaks against the interpretation as a landslide mega-block (**Plate 1**) since, if the inlier was a landslide, the northern contact (i.e. the base of the "landslide") would most likely be parallel to the overall bedding (since a landslide would fall to rest on the overall bedding surface). It may here be noted that the presence of the fault along the northern margin of the meta-psammite will not invalidate the above argument on lack of parallelism, since the movements on the fault have been so minor. It is, of cource, theoretically possible that syn-sedimentary bedding-rotation and associated erosion had taken place prior to "landslide" deposition, so that bedding was inclined compared to the "original" subaerial horizontal sediment surface, causing the base of a "landslide" *not* to be parallel to bedding. However, this is unlikely, since the bedding has a constant orientation around the whole inlier. The lack of parallelism is therefore an argument against a landslide model for the inlier. Furthermore, if the inlier was a landslide block, the structural elements in the block would most certainly have been rotated compared to the areas outside the HDM, and not be parallel to the outside elements as in the present case. In addition, the degree of brecciation is very low, and probably too low to be caused by "shaking" related to landsliding. All these features implie that the inlier cannot be interpreted as a landslide block.

(2): Tight anticline: The hypothesis that the Osen meta-psammite (and the TSW) represents the core of an anticline must be rejected. Since the orientation of bedding is the same on all sides of the inlier, an anticlinal fold would again have to be of an overturned isoclinal type with completely transposed limbs to fit the bedding data. However, the conglomerates in the area contain no signs of axial planar fabrics or any other possibly related deformation. It must be considered very unlikely that such a strong deformation as isoclinal folding could take place without leaving any signs of deformation in the conglomerates. It is important to notice that the presence of opposite grading in different sandstone lenses does not imply the presence of opposing limbs in a fold structure, since the lenses may display both normal and reverse grading.

(3): Imbrication / reverse faulting: Some degree of reverse movements has taken place along the fault at the northern margin of the Osen inlier. However, as shown in Sect. 5.5.7, the movements are only minor, and have only limited importance for the presence of the inlier.

5.4.4 THE LITLETEIGEN GREENSCHIST INLIER

5.4.4.1 DESCRIPTION

The inlier

The Litleteigen greenschist inlier is located in the steep mountain cliffs and slopes just to the west of the southern-most end of the Vassetvatnet lake, on the southwestern side of the Strupeneset headland (**Plate 1**). The name of the inlier is taken from the economical 1:5000 map "Svardal". Although the outcrop is mainly surrounded by the Devonian conglomerates, it appears to be bounded by the lake Vassetvatnet towards the east. The outcrop may therefore not be an inlier *sensu stricto*. However, since Devonian rocks are also present all the way on the southeastern side of the lake, i.e. at the lake shore just on the opposite side of the outcrop, and the distance between Strupeneset and the Devonian rocks to the south-southeast of Strupeneset is only **150 m**, the outcrop is treated as an inlier. The outcrop is essentially exposed in a high N-S oriented cliff that forms the western termination of the inlier, and stretches northwards from the lake (**Plate 1**). The cliff is partly vertical and partly overhanging, and is in the following treatment termed the "western vertical cliff". On the main map (**Plate 1**), the whole inlier is located to the east of the "western vertical cliff", in an area forming a slope rising steeply towards the NW from the lake Vassetvatnet. This slope is in the following termed the "hillside slope". The outcrop may be studied along the base of the "western vertical cliff". The base of the "western vertcal cliff" itself follows the "hillside slope" as the slope rises steeply from the lake Vassetvatnet. From the "western vertical cliff" and eastwards, the "hillside slope" itself is generally covered by a Quaternary talus-cone with a dip close to the angle of repose. The outcrop is best exposed and may be best studied at the base of the "western vertical cliff" at an elevation of about **120–130 m** above sea level, i.e. as far up the "hillside slope" as it is accessible. In the few and very small exposures located along the "western vertical cliff"-base from the road and halfway up the hillside, the inlier consists of meta-psammitic and meta-semipelitic rocks. At the exposure situated highest on the "hillside slope" at the "western vertical cliff" base, a greenschist and possibly a serpentinitic rock, as well as meta-greywacke with minor amounts of green-grey phyllonites, are present. Although most of the exposed parts of the inlier occur in the inaccessible "western vertical cliff", some information may be obtained on the rock types there. Viewed from a distance, the rocks there look grey and green, and it is likely that they also consist of greenschist and/or meta-greywacke. Whitish-coloured vertical water-flow traces are “painted” on the cliff, suggesting that carbonate might be present in the upper parts. To the east of the cliff, the inlier is very poorly exposed due to the extensive cover of the Quaternary talus cone. The rocks present in the Litleteigen inlier are very common in the substrate around the HDM. Brecciation has not been observed in the very limited outcrops that are accessible.

Contacts

The western contact of the inlier is located high up into the N-S oriented "western vertical cliff" and is thus inaccessible. With its elevated position in the cliff, the contact may be termed the "upper contact". When viewed from a distance, the position of the contact is seen from the change of rock colour from the dark-coloured inlier to the light-coloured Devonian conglomerates. The contact surface appears to be very irregular, and is therefore interpreted as a primary depositional unconformity. It is roughly oriented with a steep dip towards the E or SE.

At the base of the "western vertical cliff", in the area located in the very *uppermost* parts of the northwestward rising "hillside slope", the exposure of substrate rocks in the cliff does not immediately disappear below Quaternary deposits at the base of the cliff, but continues for **some metres** from the cliff and out on the "hillside slope", forming an accessible short belt of substrate rocks along the cliff. Still further towards the east, the belt disappears below Devonian rocks which are also present on the "hillside slope" in the slope's uppermost parts, near the cliff. The local contact between this substrate belt and the Devonian rocks to the east was not observed.

The northern contact (which may also be termed the "lower contact" since it is not located in the "western vertical cliff", but at a lower altitude below the "western vertical cliff"), has not been observed, since it appears to be covered by the Quaternary talus cone on the "hillside slope". The position of the contact indicated on the main map (**Plate 1**) is based on the local topographical expression, i.e. the contact is drawn to follow the topographic "line" representing the change from the moderately dipping "hillside slope" to the subvertical Devonian cliffs. It can therefore not be completely ruled out that the contact is actually located nearer to the "western vertical cliff", which would imply that it could occupy a more "northeastern" or even "eastern" position.

A fault with orientation **124/61 S** is present along the base of the N-S oriented "western vertical cliff". The fault does not define the lower (i.e. northeastern) contact of the inlier substrate rocks, since the substrate rocks are present also below (northeast of) this fault. This relationship between the substrate and the fault can for example be observed at the substrate belt described above (located beside the cliff-base in the uppermost part of the "hillside slope"). It is, however, very likely that the lower (i.e. "northern/northeastern") contact is also a primary depositional unconformity, since several unconformity exposures are present nearby.

On the eastern side of the Vassetvatnet lake, across from the inlier, no intra-Devonian substrate outcrops are present, and the Devonian sediments are found to form continuous exposures.

Adjacent Devonian sediments

Devonian sediments situated immediately adjacent to the contacts of this inlier have not been observed. Devonian conglomerates can, however, be observed in the road-cuts to the north (at the entrance of the road tunnel) and south of the inlier, but these rocks are not more deformed than elsewhere in the HDM, indicating that the situation in this respect corresponds to the Høgdene and Osen inliers.

The contact-bedding relationship

The contact surface between the Devonian bedding and the substrate has not been observed, but appears to be oriented with an essentially N-S strike and a steep E dip. If this is correct it means that the WSW-ENE striking, and steeply SSE dipping bedding in the area approaches the contact at a rather high angle.

5.4.4.2 INTERPRETATION

The Litleteigen inlier is interpreted to be an outcrop of "rooted" substrate that forms a sub-Devonian topographic spur, i.e. implying that erosion has revealed the substrate below the Devonian cover, and that the contact around the inlier is essentially a primary depositional unconformity. The fault present at the base of the cliff is interpreted to be a "late" feature cutting through *both* the Devonian and the substrate inlier.

Sub-Devonian high below Teigafjell: The fact that the substrate crops out in a) the Litleteigen inlier; b) the Teigafjellet Substrate Wedge; and c) the valley going up between the Stigen Devonian spur and Nonsnova, has important implications. Firstly, it shows that the cover of Devonian sediments forms only a thin carpet on top of the substrate in the eastern part of the study area. Secondly, since several primary depositional unconformities have been found near by (see **Plate 1**), it must be concluded that a "substrate-high" is probably present below the Devonian cover from Teigafjell to Leirvågfjell. This "high" may be considered as a "palaeo-mountain", which may be denoted the "sub-Devonian-Teigafjell/Leirvågfjell". Eastward from this area , the sub-Devonian "high" forms a "slope" or "mountainside" from the top of the "Sub-Devonian-Teigafjell/Leirvågfjell" down towards the Vassetvatnet lake/Storelva river. The Litleteigen inlier crops out in this "mountainside", and the Devonian conglomerates of the Stigen and Strupeneset Spurs are situated on this eastward-dipping Palaeo-mountainside. The present-day erosional level of the substrate in this mountainside corresponds roughly to the surface that was exposed when the deposition of the Devonian sediments started.

As for the Høgdene and Osen inliers, the interpretation above is also partly based on the rejection of alternative models. Since the particular reasons for rejecting the alternative models may vary from inlier to inlier, the reasons related to the Litleteigen Inlier are discussed below.

(1): Landslide: It is considered very unlikely that the Litleteigen inlier is a landslide. The main reason for this is the presence of several unconformities between the Devonian rocks and presence of proper-substrate rocks only **some hundred metres** away (at the end of the Vassetvatnet lake about **300 m** to the east of the inlier; at Nonsnova about **800 m** to the north of the inlier; and at the Teigafjellet Substrate Wedge about **900 m** to the northeast of the inlier), as well as the presence of near-by "proper" substrate, such as in the Svardal area only **300 m** to the southeast of the inlier. These unconformities indicate that the cover of Devonian sediments in the hillside from Teigafjell to Storelva is very thin, and the inlier is therefore most likely a sub-Devonian topographic feature exhumed by erosion, and not a landslide.

(2): Tight anticline: Although bedding measurements are not present close to the inlier, it is reasonable to assume that the constant bedding orientations at Teigafjell, some **800 m** to the northwest, is also present around the Litleteigen inlier. An anticlinal fold would therefore have to be of an overturned isoclinal type with completely transposed limbs to fit the bedding data. The hypothesis that the Litleteigen inlier represents the core of an anticline must be rejected on the basis of similar features and arguments as those related to the Høgdene and Osen inliers.

(3): Imbrication of "basement-cover" unit (high angle reverse faulting): The fault along the base of the "western vertical cliff" cuts *through* the substrate inlier, and this fault therefore cannot be used to postulate that imbrication or reverse faulting is responsible for the presence of the inlier, since this would require that the fault go at the base of the inlier. Measurements of the semiductile to brittle shear fabric show that the fault is essentially planar, thus making listric imbrication unlikely within the scale of observation. The fault can be observed to continue on the eastern side of the Vassetvatnet lake/Storelva river (**Plate 1**), and no other faults that could have lifted up the outcrop are present in the area. The amount of movement on the fault is uncertain, but is probably minor, since substrate rocks are present on both sides, and fault fabrics are modest. The direction of movement is uncertain.

(4) Strike-slip/oblique-slip faulting: Since the movement direction on the fault along the base of the "western vertical cliff" is uncertain, it is possible that the movements on the fault were more strike-slip or oblique-slip, than dip-slip. Anyway, such faulting alone cannot explain the presence of the inlier, since the substrate rocks are present on both sides of the described fault.

5.4.5 THE KVANGAGJELET SYENITE

5.4.5.1 DESCRIPTION

General

The Kvangagjelet syenite is located at the southern margin of the HDM (**Plate 1**), and is transected by the western profile (**Plate 2**). "Kvangagjelet" is the name of the quite prominent canyon following along the southern margin of the syenite, at its western part. The name is in common use by the local population, but has not been included on presently published topographical or economical maps. This lack of names on official maps of the area has made it necessary to use the name "Kvangagjelet" on the maps of the present work. The Kvangagjelet syenite is different from the intra-Devonian substrate outcrops described above in that the syenite is not an inlier, since it is not "surrounded" by younger (Devonian) rocks (**Plate 1**). In spite of this, the syenite could theoretically occupy an "intra-Devonian" position, and it is therefore treated here.

Devonian sediments have not been observed along the accessible parts of the southern margin of the syenite (although it is likely that the Devonian sediments covered the ares to the south of it; see discussion in Ch. 6), and the syenite itself is in fault contact with the rocks of the Høydalsfjorden Complex to the south. The only way in which the syenite could be intra-Devonian in this situation would be that the syenite is a Devonian landslide block present within the strata of the basin, i.e. with Devonian sediments beneath, and that this "sediment-syenite-sediment package" is now in fault contact with the rocks of the Høydalsfjorden Complex. (Although more unlikely, a land slide could also fall to rest directly on a HC rock surface, i.e. without Devonian sediments beneath).

The syenite

The rock body is at least **1.5 km** long; **125 m** wide in the map plane; and up to **250 m** thick as measured on the mountain side going down into the Kvangagjelet canyon. The syenite is generally massive, with an essentially "magmatic" texture preserved as judged from the outcrops. Along the southern margin, the syenite locally has a vague foliation with a WNW-ESE to NW-SE strike and a dip of about **60° NNE to NE**. Plutonic analogues to the present rock have not been found in the substrate elsewhere in the region. The nearest larger plutonic body in the Lower Palaeozoic part of the substrate, is the Bremanger Granodiorite situated **35 km** to the northwest. The Kvangagjelet syenite has not been radiometrically dated, and its relationship to the Høydalsfjorden Complex is uncertain. It is therefore at present an open question whether the syenite is Caledonian or Precambrian. As mentioned in Sect. 4.2.2.4, Kolderup (1925, 1928) stated that the Kvangagjelet syenite to him looked similar to the syenites that forms the substrate to the Kvamshesten Devonian deposits. In later works, these syenites at Kvamshesten have been assigned to the Precambrian Dalsfjord Suite (/Complex), which Corfu and Andersen (2002) dated to **1634 +/-3 Ma** (age of intrusion). This opens for the possibility that the Kvangagjelet syenite is Precambrian. However, the Lower Palaeozoic rocks of outer Sunnfjord and Sogn contain numerous large bodies of intrusive rocks, and a (late?) Caledonian age therefore cannot be excluded for the syenite. The Kvangagjelet syenite shows signs of brecciation in places, although this may be difficult to observe at the outcrop due to the general lack of small-scale markers (like e.g. foliation, lithologic banding, etc.) in the massive rock, and due to cover of lichon. It is uncertain whether the brecciation is fault-related or regolithic, but the amount of brecciation appears to increase towards the fault in the canyon of Kvangagjelet, suggesting a possible connection with movements along this fault.

Contacts

The northern contact towards the Devonian deposits is exposed at one locality (at the river from Håsteinsvatnet). There, the contact is a **100 %** primary depositional unconformity (**Fig. 5.42**, **5.43**). The southern contact is defined by the fault in the canyon of Kvangagjelet. The amount of displacement on this fault is uncertain, but is probably small (see **Plate 2**, western profile). The displacements probably have a significant dip-slip component since the fault trace is curved in map view. Devonian sediments do not appear to be present on the southern side of the fault.

Adjacent Devonian sediments

The Devonian sediments at the northern margin of the syenite show no signs of tectonic deformation. The conglomeratic clasts are, from the conglomerate/syenite contact and **hundreds of metres** northwards, composed almost exclusively of a similar syenitic rock.

The contact-bedding relationship

The angle between the bedding and the contact can only be observed at the upper (northern) unconformable contact. The angle is here small, i.e. bedding is subparallel to the contact. The contact and bedding dip about **60° NNE**, as does also the foliation in the syenite, where it is developed. As implicit on the western profile (**Plate 2**), the angle between the Devonian bedding and the original irregular syenite palaeo-surface (stippled air line) across the syenite might have been high, as is also the case for the Høgdene, Osen and Litleteigen inliers.

5.4.5.2 INTERPRETATION

The interpretation of this outcrop is not straight forward. The uncertainty arises because of the presence of the fault along the southern margin of the syenite and because of the lack of Devonian sediments to the south. The syenite is either (1) an integral part of the Devonian succession (i.e. a landslide block), or (2) an integral part of the substrate, and these two alternatives are discussed below.

(1) Integral part of the HDM, i.e. landslide block. It is theoretically possible that the syenite is part of a landslide deposited in the Devonian basin. This would generally require Devonian sediments to be present also *below* the syenite. A combined rock "package" would then arise, with Devonian sediments at the bottom, the syenite body above, and finally a cover of sediments (now partly eroded away) on top. This “package” would then, at present, be in fault contact with the rocks of the Høydalsfjorden Complex to the south. This would be the only way that the syenite could be intra-Devonian in the present situation, since Devonian sediments are absent on the southern side of the syenite. The clasts in the Devonian conglomerate in the area are identical to the syenite, and it is therefore possible that both the syenite and the clasts could have been "deposited" from similar source rocks. The brecciation in the syenite is difficult to interprete, and it can therefore not be excluded that it is related to "shock"-effects during rock-sliding. Sand dykes, which may be common in landslides, have not been observed, although this absence could be related to i) lack of sand in the subjacent layers, ii) large depth to such layers, or iii) unfavorable conditions for formation of sand dykes. Apophyses from the syenite into the substrate HC rocks have not been observed, and this may suggest that the syenite is not intrusive into the substrate, although it may still be part of the substrate as a mega-olistolith in a possible melange.

(2): Integral part of the substrate: If the syenite is an integral part of the substrate, the syenite may be:

(a) part of the Precambrian basement (as suggested by Kolderup 1925, 1928)

(b) intrusive into the Lower Palaeozoic substrate

(c) a mega-olistolith in a Lower Palaeozoic melange

The present section focuses on the question of whether the syenite is ***at all*** part of the substrate, or not. Hense, extensive discussion of which of these sub-alternatives that are the most likely, are not carried out.

No Devonian sediments are present on the southern side of the syenite, possibly suggesting that the syenite may be part of the substrate. Furthermore, none of the substrate inliers described earlier in the chapter are landslides, and this may be taken to indicate that the syenite is not a landslide. Apophyses from the syenite into the substrate rocks have not been observed, and this may suggest that the syenite is not intrusive into the substrate, although it could still be a mega-olistolith in the Lower Palaeozoic substrate. The alternative that the syenite is part of the local Precambrian basement, as a 'basement window', is not likely. This is due to the fact that the syenite body is largely unfoliated, with its magmatic texture preserved, whereas the local Precambrian rocks are mylonites constituting the Nordfjord–Sogn Detachment Zone.

Evaluation of the two models: It is difficult to test the landslide model, since it is unknown whether Devonian sediments are actually present below the syenite. The presence of the fault along the southern margin of the syenite makes it difficult to test whether the syenite is an integral part of the substrate or not, since the original contact relations to the surrounding substrate is obscured. From the above discussion, it appears that the body is either an integral part of the HC – or a landslide within the Devonian deposits. As shown above, arguments can be raised in favour of both these models.

An argument in support of the landslide model could be that the syenite body and above-lying Devonian sediments (Blåfjell Fm), are 'exotic' to the rocks of the Høydalsfjorden Complex (Sect. 5.2.2.1), i.e. do not exist in the HC. The composition of the Kvangagjelet syenite and the above-lying Devonian clasts are identical. This could indicate that they were both deposited from a common source area. The syenite is positioned only **~1 km** north of the E-W trending NSDZ (Standalen segment), which represents the southern margin of the HC, and which also probably defined the southern margin of the Håsteinen Basin during deposition of the Devonian sediments. High relieff probably existed along the basin margin during deposition, and the syenite could then have moved as a landslide into the basin.

Nevertheless, caution must be executed in the present case, when using "exotic" rocks as an argument. The reason is that the Kvangagjelet syenite could theoretically be an integral part of the substrate, and still be covered with Devonian sediments of identical composition. This has to do with the relationship between Upper Plate rocks (basin floor) and Lower Plate rocks (source area): any rock type present in the Lower Plate, could in principle also be be present in the Upper Plate, and vice-versa. This can be illustrated as follows: in the thesis area, the HC rocks define the Upper Plate that has the Håsteinen deposits on top. At Bremangerlandet west of the Hornelen Devonian Massif, a Caledonian nappe stack of Caledonian rocks and Precambrian gneisses constitute the Upper Plate that forms the "floor" to the western part of the Hornelen Devonian Massif. At the time when the basins formed, Upper Plate (basin floor) -type rocks were also present *outside* the marginal faults of Devonian basins, defining Lower Plate *source rocks* for the Devonian sediments. During the long time period that the rocks of the Caledonian nappe stack were still present in the source areas, the sediments in the Devonian basins would be of similar type. When the basin-controlling faults formed, some rocks of the Caledonian nappe stack would accidentally be *within* the basin, forming the Upper Plate, and some would be *outside*, forming the Lower Plate. This means that if the faults had formed at a different position, any source rock now supplying sediments into the basin, could instead have become the basin floor inside the basin. The fact that syenitic clasts were deposited into the Håsteinen basin shows that this rock type was present in the source area. This in turn means that the rock type, in principle, could also have become present in the Håsteien basin floor. Hense, the Kvangagjelet syenite could be part of the substrate, and still be covered by Devonian sediments of identical composition.

At the present stage the available data do not seem to clearly support one model at the expence of the other, although the exotic character of the Kvangagjelet syenite and the above-lying Devonian sediments, in relation to the substrate, could possibly be taken to indicate that the landslide model is a bit more likely.

In addition to the models considered above, the following rejected models shall be briefly mentioned:

(3): Tight anticline: Due to the lack of Devonian sediments on the southern side of the syenite, it is not possible to test whether the body is present within the core of an anticline. However, since none of the other inliers are anticline-cores, it is very unlikely that the syenite represents such a core.

(4): Imbrication: Due to i) the presence of a fault along the southern margin of the syenite, ii) the lack of Devonian sediments on the southern side of the fault, and iii) the fact that it is completely uncertain whether Devonian sediments are present below the syenite, it is not possible to test whether the syenite is imbricated. Based on the situation for the other inliers, which are not imbricated, an imbrication must be considered unlikely also for the syenite.

5.4.6 SUMMARY

The Høgdene meta-psammite inlier, the Osen meta-psammite inlier, and the Litleteigen greenschist inlier are interpreted as sub-Devonian topographic highs of "rooted" substrate that have been revealed by erosion. The Kvangagjelet syenite could belong to the substrate, although it is possibly a bit more likely that it is a landslide within the HDM.

5.5 STRUCTURAL AND METAMORPHIC DEVELOPMENT OF THE HDM

5.5.1 INTRODUCTION

General

Section 5.5 deals with the structural and metamorphic development of the HDM. The HDM has been subjected to one phase of semiductile/ductile deformation, denoted D_1, which has folded the massif into a sharp (angular), open F_1-syncline termed the Osstrupen syncline (**Plate 1**, **Plate 2**), and local parasitic F_1-folds with axial planar S_1-cleavage. The accompanying low-grade metamorphism is termed M_1. This D_1-deformation is the dominating deformational event in the HDM. Post D_1-deformation is only present in the form of later shear zones and faults. It is possible that parts of the deformation occurred prior to complete consolidation of the sediments, but conclusive evidence for such soft sediment deformation has not been found. The subject of "soft sediment deformation", together with a general discussion on folding mechanisms for the Osstrupen syncline, are treated in Ch. 6 as part of the final discussion on the deformation of the HDM.

The current section (5.5) starts with a structural analysis of the Osstrupen F_1-syncline (Sect. 5.5.2), including a comparison with the local substrate structures as well as with the Grøndalen Syncline of the Hornelen Devonian Massif. This is followed by structural analyses of the parasitic F_1-folds at Gravaneset, and the related axial planar S_1-cleavage (Sect. 5.5.3), including a comparison with the deformation in the locally underlying substrate rocks. Thereafter, the structural features of Gravaneset are compared with data on tectonomagnetic fabrics from Torsvik et al. (1987) (Sect. 5.5.4), and subsequently, the stratigraphic thickness of the HDM is discussed (Sect. 5.5.5). This is succeeded by a discussion of the M_1-metamorphism (Sect. 5.5.6), and thereafter the faults cutting the HDM are described (Sect. 5.5.7). Finally, the three-dimensional architecture of the HDM is discussed (Sect. 5.5.8), followed by a summary of the structural and metamorphic development of the HDM (Sect. 5.5.9).

Due to the large scale of the folds in the HDM, fold axes cannot be measured directly in the field. In the present work, the fold axes are therefore obtained by construction of stereographic π-axes found from π-circles based on poles to bedding planes. The average values of pole populations have been estimated from contoured plots with down to **1 %** contour interval, giving very presice maxima. (On figures, however, the plots are *shown* with **5 %** and **2 %** contour interval to ensure legibility). These stereographic maxima have been used for *visual* construction of π-circles yielding the mean value of the fold axes.

5.5.2 THE OSSTRUPEN F_1-SYNCLINE

5.5.2.1 INTRODUCTION

Section 5.5.2 deals with the geometry of the Osstrupen syncline and analyses the structural elements by stereographic means. The significance of this syncline for the important general 3-D architecture of the HDM is described later (Sect. 5.5.8).

As can be seen on **Fig. 5.50** and **Plate 1**, the major massif-wide Osstrupen syncline has an axial trace trending WNW–ESE, which goes through the centre of the massif. The syncline is an upright, subcylindrical, open, plane fold with a vertical axial surface, straight limbs, and a fold axis dipping **53°** towards the ESE. A narrow hinge zone is apparently present only on the northern side of the axial trace (Western profile, **Plate 2**).

The significance of sandstone lenses for the establishment of the Osstrupen syncline

The basis for the establishment of the synclinal structure is the occasional presence of sandstone lenses which can be found scattered throughout the generally massive and structureless proximal alluvial fan/debris flow deposits (Sect. 5.2.2.1). The sandstone lenses are frequently located far apart, and since the debris flow beds themselves are often difficult to discern, none or only vague bedding indications are present between the lenses. Therefore, a particulare stratigraphic level cannot be traced accurately in the field. The orientation of these isolated lenses is, however, very constant over large areas, and they are considered to be reliable as bedding indicators (Sect. 5.2.2.1). The lenses frequently contain parallel and planar lamination of millimetre to centimetre scale, yielding precise and consistent bedding recordings. The sandstone lenses are interpreted to be remnants of thin sand deposits which locally may have formed on top of the debris flow deposits, due to fluvial processes at the waning stage of the flows, or shortly after (Sect. 5.2.2.1).

Primary depositional angle of alluvial fans

In the following discussions on structural aspects of the Osstrupen syncline, the quite strongly inclined orientation of the present-day bedding in the HDM will be used to indicate the amount of tectonic rotation (due to folding and other factors) that effected the Devonian layers after deposition. However, when discussing such rotations, it is important to note the following: the fact that the sandstone lenses are present in alluvial fan sequences does not imply that the lenses were deposited on strictly horizontal surfaces. Generally, an average *primary inclination* of up to **10°** from the horizontal might be expected for surfaces of such debris flow deposits at the stage of deposition (e.g Blatt et al. 1980; Collinson 1978; Hooke 1967). However, since the actual deviation from the horizontal cannot be presicely estimated, no correction is made for this source of error. The very high consistency in the orientation of bedding, and the very low angle of primary deviation from the horizontal (max **10°**), make the error (in assuming an original horizontal bedding) small enough to be neglected for the structural topics under consideration here.

The primary overall *strike* of bedding may have had a larger variation in magnitude than the dip, at the stage of deposition. This is due to the gently curving (convex upward) and shallowly dipping overall surface of alluvial fans at this stage. However, when these bedding surfaces of alluvial fans are rotated to a steeply dipping position by e.g. folding, the strike variation will be strongly "reduced", (to eventually become comparatively small) as it can be seen to be in the HDM.

Terminology and names

For the description of the Osstrupen syncline, the following subdivisions have been made (c.f. **Fig. 5.50**):

"The northern fold-half": Denotes the area to the north of the axial trace. The northern fold-half is furthermore divided into two subareas:

"The northern limb": Denotes the outer and major part of the fold-half. The limb is straight and contains constant bedding orientations.

"The hinge zone": Denotes the narrow belt, along the axial trace, with gradually changing orientation of bedding towards the fold axis (axial plane).

"The southern fold-half": Denotes the area to the south of the axial trace. The southern fold-half is identical with:

"The southern limb": Denotes the whole area to the south of the axial trace, as no hinge zone is present in the southern fold-half.

In map view, the northern fold-half covers an area about the same size as the southern fold-half (**Plate 1**).

Distribution of formations and facies in the limbs

Most of the northern limb consists of the coarse debris flow conglomerates and breccias of the Vikafjell Formation which is present everywhere except for the marginal parts in the west and northwest, where the sandstones of the Gravaneset, Mannen, Stegabruna and "Galmannsskåra" sandstone units occur. In the western areas of the HDM, the Vikafjell Formation continues across the axial trace into the southern limb (Sect. 5.2).

Most of the southern limb consists of coarse debris flow conglomerates and breccias of the Blåfjell Formation. The formation is present from Fjellsenden in the east and all the way to the areas around Blåfjellnipa in the west. The areas from around Blåfjellnipa and westwards / northwestwards consist of coarse conglomerates and breccias of the Vikafjell Formation. The Vikaholmen and Novene sandstone units, in the westernmost parts, are the only sandstone deposits present in the southern limb (Sect.5.2).

The Vikafjell and Blåfjell Formations are lateral eqivalents and grade into each other in the west.

5.5.2.2 THE FOLD ORIENTATION

From the main map (**Plate 1**) and the detailed maps (**Appendix A**), it may be seen that the orientation of bedding within each of the *limbs* of the Osstrupen syncline is remarkably constant. Furthermore, although impressive as a document of Devonian deformation, the Osstrupen syncline is a fairly simple structure. It is therefore unnessecary to subdivide the *limbs* into smaller subareas with separate stereo-plots of each area. Separate stereo-plots are, however, presented from two very important *fold closure areas*, i.e. at Nonsnipa (near Osen) and to the south of Gravanesholten.

Orientation of bedding in the northern limb

The northern limb constitutes most of the northern "fold-half". From a stereo-plot of the **376 poles** to bedding recordings throughout the HDM, and a contoured diagram of these pole plots **(Fig. 5.51a** and **5.51b**), a precise visual estimate yields an average orientation of bedding in the northern limb of **070/62 SE**. In these plots, the poles to the bedding in the northern limb define the "northernmost" part of the pole girdle containing the particularly high pole concentration. Most bedding orientations deviate less than **10–15°** from the average orientation. As seen on the main map (**Plate 1**), there are no gradual or systematic changes in strike or dip of bedding along the northern limb from the western to the eastern areas, and the use of a single plot for all the limb bedding poles is appropriate. The remarkably constant orientation of bedding within each limb is characteristic for both limbs in the Osstrupen syncline.

<u>*Areas with deviating bedding orientations*</u>: Although the orientation of bedding is generally constant, the main map (**Plate 1**) shows that two minor areas within the *outer* part of the northern limb area, contain bedding orientations which deviate more than **10–15°** from the average trend.

The first area is located in the easternmost to northeasternmost area of the summit of Vikanipa, close to the contact between the Devonian rocks and the substrate. A few bedding-strike measurements deviate consistently from the general trend, and thereby define a local anomalous sub-area. The *dip* of bedding is unchanged as compared to the rest of the limb. The bedding recordings have *strike* directions around **090°**, and at the easternmost tip of Vikanipa (**Plate 1**) even **105°**, which is **35°** from the average of **070°**. However, bedding with orientation around the average of **070/62 SE** is present both in the Vassetbrekka area **350 m** to the south of the summit of Vikanipa, and also about **400 m** to the west of the summit. This shows that the deviating area is surrounded by an area with average bedding orientations. The deviation is considered to be merely a local irregularity of either primary sedimentary or tectonic origin.

The other area with deviating bedding orientations is confined to the Stigen Devonian spur. In this area, some of the measurements give strike orientations within the usual **10–15°** deviation from the average of **070/62 SE**, but the majority of the measurements exhibit strike orientations from **090°** to **106°**, which deviate **20–36°** from the average strike of **070°**. The deviating bedding orientations appear to be mixed with bedding orientations lying close to the **+/- 10-15°** from the average. This indicates that no systematic change in strike of bedding is present along the spur. The deviation of bedding is therefore most likely due to a variation in primary orientation which is larger than usual. In the stereograms (**Fig. 5.51a** and **5.51b**), the slightly larger scatter of the bedding poles of the northern limb compared to the southern limb is largely due to these deviations.

Bedding constancy across faults: Several prominent faults with map traces oriented from E-W to WNW-ESE transect the northern limb of the Osstrupen syncline (Sect. 5.5.7). These include the two faults on Vikafjell, the fault on Leirvågfjellet, and the fault along the northern margin of the Osen inlier between Osen and Teigafjell (**Plate 1**). The faults are further described in Sect. 5.5.7. As mentioned above, the average bedding orientation is constant in the northern limb, and this is also valid across these faults all the way from Vikanipa in the north to Leirvågfjellet, Nonsnova and Teigafjell in the southeast (**Plate 1**). The fact that the bedding does not change orientation across the faults shows that rotation of individual fault blocks between the faults has not occurred. The displacements on these faults appear to be small (Sect. 5.5.7).

Orientation of bedding in the hinge zone

The hinge zone in the northern fold-half is located along the axial trace to the south of the northern limb area (**Fig. 5.50** and **Plate 1**). The boundary between the northern limb and the hinge zone may be roughly drawn from Straumsnes and eastwards along the fjord Osstrupen to the areas around Osen (**Fig. 5.50** and **Plate 1**). The width of the hinge zone in map view is at least **500–600 m** in the westernmost areas, and possibly **400–500 m** in the Osen area. A possible further continuation eastwards along the southern side of the Osen inlier, towards the village of Svardal, is more uncertain, as the hinge zone appears to wedge out towards the east. The presence of a slightly longer and wider fold arc in the western part than in the eastern part of the Osstrupen fold is analogous to the situation in the Grøndalen Syncline along the southern margin of the Hornelen Devonian Massif (Bryhni & Lutro 1991b, see comparison below, Sect. 5.5.2.6).

In the hinge zone, the strike orientation of bedding changes *gradually* along strike from the hinge line to the limb segment, as opposed to the *straight* bedding in the limb. This is well illustrated in the area located far west, between Straumsnes and the axial trace to the south of Gravanesholten (**Plate 1**). The strike orientations of bedding are, when followed from north to south, gradually rotating anticlockwise from the typical ENE-WSW to an orientation of NNW-SSE in the southern limb, and in the stereoplots (**Fig. 5.51a** and **5.5.1b**), the poles are distributed along the girdle between the north limb and south limb maxima.

<u>Faults in the presice fold closure in the west</u>: In the central parts of the plateau stretching southwards from Gravanesholten towards Blåfjellnipa, a fault follows almost exactly along the axial trace (**Plate 4**). The age relationship between the fault and the D_1-deformation is uncertain, although the fault is probably post-D_1 (Sect. 5.5.7). In the *western* part of the plateau, the strike orientation of bedding appears to change abruptly instead of gradually across this fault, i.e. from the northern hinge zone to the southern limb (**Plate 4**). It is, however, considered very unlikely that the fault alone has produced this abrupt change of bedding, particularly since bedding with more gradual and continuous change of orientation is present **some tens of metres** to the *east* along the fault. The amount and direction of displacement on the fault is uncertain.

The most likely explanation for the abrupt change of bedding orientation on the western part of the plateau, is that the change was essentially produced during the folding that formed the Osstrupen F_1-Syncline in the HDM. Space problems occurred in the fold closure during this deformation, and this possibly caused the folding to be accompanied by "movements" along the axial plane zone. These movements probably took place in the form of reconstitution of the positions of individual clasts, and thereby produced a"weakened zone" along the axial trace, as well as the abrupt change of bedding orientation across it (see Ch. 6 for a discussion on folding mechanisms for the Osstrupen syncline). The present fault, considered as a post-D_1 feature, later formed in this weakened zone, and therefore need not be responsible for the abrupt change. The Grøndalen Syncline in the Hornelen Devonian Massif (see Sect. 5.5.2.6) show features that are analogous to the Osstrupen fold: the Grøndalen fold closure has a similar abrupt change of bedding orientation from one limb to the other, and also has a fault along the axial trace zone.

Orientation of bedding in the southern limb

In contrast to the northern fold-half, the southern fold-half does not appear to have a hinge zone. Instead, the fold-half contains only constant bedding orientations, and the whole area is thus interpreted to constitute the southern *limb* (**Plate 1**).

From the contoured stereoplot of the **376 bedding poles** throughout the HDM, where the southern limb is defined by the west-southwesternmost high pole concentration (**Fig. 5.51a** and **5.51b**), a precise visual estimate yields an average bedding orientation of **161/62 NE** for the southern limb. Most poles deviate less than **10–15°** from this average orientation. The bedding has this average strike orientation along practically the whole southern limb, and there is no signs of a systematic or gradual change in orientation along the limb.

The bedding orientation along the axial trace of the southern limb will now be described in some more detail. In the western parts, the absence of a hinge zone is clearly seen, as the average constant bedding orientations can be traced all the way into the precise fold closure. At Osen, however, the exact fold closure (or hinge point) is not exposed (due to grass/vegetation). When looking southwards from the Osen fold closure, the first rock exposures appear about **200 m** from the inferred position of the axial trace. In these first exposures, the strike direction of bedding is similar to the average bedding-strike direction of the limb. Theoretically this may mean that a hinge zone, with gradually changing bedding strikes, could be present in the **200 m** wide zone.

However, since no indications of a change of bedding strike is present at the first exposures, it is considered most likely that the hinge zone is absent also in the Osen area, as in the areas further to the west.

It is thus concluded that the constant bedding orientations in the southern limb continue all the way into the axial trace.

Area of deviating bedding orientation: Although the bedding orientations along the southern limb is generally remarkably constant from the western to the eastern areas, bedding orientations that deviate more than **10–15°** from the average have been recorded in one single area: the small Neset penninsula at the southern shore of lake Svardalsvatnet. In this area, the bedding *strike* does not deviate from the average, whilst the *dip* appears to be **30–34°**, i.e. shallower than the average dip of **62°**. However, bedding that dip with the normal magnitude are present both further south along strike, and at Nonsnipa to the west, and the Neset area is thus surrounded by areas with normal bedding orientation. The deviation is considered to be merely a result of a local variation in the primary orientation that is larger than usual, and does not appear to indicate a systematic change in the bedding orientation, which is otherwise very constant along the rest of the limb.

Bedding constancy across faults: On the southern limb, only the fault going from Breiskåra to Østre Håsteinen (**Plate 1**) is topographically prominent. Based on the bedding measurements around the lake of Håsteinsvatnet, the orientation of bedding does not appear to have been affected by the fault.

The orientation of the *average* fold axis for the Osstrupen syncline

The general fold axis of the Osstrupen syncline has been obtained through a stereographic π-plot. In the contoured plot showing the density distribution of the **376 poles** to bedding planes throughout the synclinal structure, the best fitting π-circle through the limb-pole maxima defines a π-axis with orientation ***115/53*** (**Fig. 5.51a** and **5.51b**), *and this defines the average fold axis for the Osstrupen syncline*.

The fact that bedding orientation on the limbs is so constant, indicates that the mean fold axis is representative for the whole massif. According to the classification suggested by Ramsay and Huber (1987), the plunge of **53 °** corresponds to a "moderately plunging" fold (**30-60°**), being near the boundary to a "steeply plunging" fold (**60–80°**).

The majority of the poles located along the very central part of the girdle (i.e. the hinge zone) are situated slightly to the west of the selected average great circle (**Fig. 5.51b**). The reason for this is that some of these poles are from the Gravaneset parasitic folds, where limbs with a NNE-SSW strike-orientation have slightly steeper bedding-dips than the rest of the Osstrupen syncline (Sect. 5.5.3.1). When selecting the great circle it is therefore most correct to make a best fit to the majority of the poles belonging to the *limbs* of the Osstrupen syncline itself. The deviation of the Gravaneset poles from the selected great circle is, however, very small, since most of the poles lie within **+/- 10-15°** from the π-circle anyway.

The orientation of fold axes in *local* fold closures of the Osstrupen syncline

The precise location of the fold closure, or the hinge points, of the Osstrupen syncline have been found in two areas along the axial trace:

The first area is situated in the westernmost parts of the HDM on the plateau which streches from Gravanesholten and about **400 m** southwards (**Plate 4**). As mentioned above, the bedding in the western parts of the plateau changes orientation abruptly instead of gradually around the fold closure, whilst a more gradual change is present in the eastern parts of the plateau (**Plate 4**). The fault running parallel to the axial trace in the western parts of this fold closure therefore cannot alone be responsible for the abrupt change of bedding orientation across the axial trace (see above). Although the exact amount of displacement is uncertain, the displacements along the fault appear to be only minor (see Sect. 5.5.7), i.e. of no importance in the following structural analyses of the local fold axis. Furthermore, the fact that bedding has a constant *dip* around the fold closure, and that the *strike* of bedding changes gradually in the hinge zone on the eastern part of the plateau, indicate that rotation of fault blocks have not taken place. Construction of π-plots to obtain the fold axis for the fold closure is therefore valid, and the best fitting π-great circle can be found from the contoured stereogram showing the density distributions of the **62 bedding poles** which define a girdle along a great circle (**Fig. 5.52a** and **5.52b**).

*The best fitting great circle yields an average fold axis with orientation **112/58** for this part of the HDM* (**Fig. 5.52b**), which is close to the general fold axis of **115/53** for the Osstrupen syncline (cf. **Fig. 5.51b**).

The other area where the fold closure has been found, is located in the central part of the Håsteinen massif, just south of Nonsnipa at Osen (**Plate 1**). The precise fold closure is not exposed, but since the locality covers only a small area, the fold closure may be accurately located by extrapolation from a large number of high quality bedding recordings nearby (**Appndix A: map no. 10a**). The **39 poles** to bedding form two distinct populations which on the contoured plot give well-defined maxima yielding a best fitting π-great circle (**Fig. 5.53a** and **5.53b**).

*The corresponding mean fold axis has the orientation **105/50*** (**Fig. 5.53b**), *defining the local fold axis in this part of the HDM.*

Although this fold axis of **105/50** deviates up to **10°** in azimuth from the general fold axis of **115/53** for the Osstrupen syncline (cf. **Fig. 5.51b**), the two axes are still fairly close. The fact that the fold axis at Osen has a more shallow dip than the axis on the plateau to the south of Gravanesholten is not the result of a gradual trend to more shallow dipping axes towards the east, but merely a result of minor local variations. This is illustrated by the fact that the limb orientations to the north and south of the Osen area are not different from the orientations further to the west.

The azimuth and plunge of the fold axes thus appear to be quite consistent and constant along the axial trace of the fold. This is indicated both by the similar orientation of the π-fold axis on the plateau to the south of Gravanesholten and at Nonsnipa to the south of Osen, and also by the generally very constant orientation of the bedding along the limbs of the fold (**Plate 1**), which will give a fold axis with essentially the same orientation also at other places along the axial trace of the fold. This is illustrated on the axial trace profile (**Plate 2**) and discussed in Sect. 5.5.8.

It is worthwile noting that this lateral *consistency and constancy* in the dip of bedding within each limb along the axial trace of the HDM, is analogous to the situation in the Hornelen Devonian Massif, where bedding has a constant dip of about **25° E** from western to eastern areas (Kolderup 1927a; Steel 1985; Seranne & Seguret 1987; Norton 1987). The situation is also analogues to the Kvamshesten Devonian Massif, where the avarage dip is about **15° E** from western to eastern areas (Skjerlie 1971), although variations can occur (Osmundsen et al. 2000). The constant dip of bedding in the HDM provides a possibility to calculate the stratigraphical thickness of the massif (see Sect. 5.5.5).

Interlimb angle

In the Osstrupen syncline the northern limb dips about **62° SE** and the southern limb about **62° NE** (**Fig. 5.51b**). In a stereogram, where a new great circle is defined by the poles to the two limb great circles, the interlimb angle will be defined by the part of the new great circle that is located between the two limb great circles. (The intersection of the two limb great circles defines the fold axis plunging **53°**). This stereographic technique yield an ***interlimb angle of 102°*** (**Fig. 5.54**).

The interlimb angle lies in the interval **70–120°**, and the fold is thus an open fold (e.g. Ramsay & Huber 1987).

Axial plane

The axial surface is defined as the surface connecting the adjacent hinge lines in a folded multilayer sequence (Ramsay & Huber 1987). The constant orientation of both the fold axes along the axial trace, and the bedding within the limbs, show that the axial surface is essentially planar, indicating that the Osstrupen syncline may be classified as a plane fold.

For the Osstrupen syncline, it is reasonable to assume that this plane closely corresponds to a plane bisecting the interlimb angle as defined by e.g. Turner & Weiss (1963) and Davis & Reynolds (1996), since both limbs appear to have experienced the same degree of deformation (rotation). The axial plane can thus be obtained through stereographic techniques (**Fig. 5.54**), *giving a mean orientation of 115/90*, i.e. exactly vertical.

This means that the fold may be classified as "upright" (e.g. Ramsay & Huber 1987). The fold axis for the Osstrupen syncline (**115/53**) is thus situated within the axial plane. The vertical attitude of the axial plane is reflected in the linear trend of the axial trace, i.e. unaffected by topographic variations (**Plate 1**).

Axial trace

The axial trace of the Osstrupen syncline has a WNW-ESE orientation through the middle of the massif (**Plate 1**), and separates the northern fold-half from the southern fold-half. As described above, the precise position of the axial trace has been found by the location of hinge points at two places; (1) on the western plateau about **400 m** to the south of Gravanesholten, and (2) to the south of Nonsnipa at Osen. Between these two areas the axial trace has been interpolated. The interpolation may be done with a high degree of confidence, since the bedding orientations in the limbs are so constant. To the east of Osen, the axial trace is positioned between the bedding recordings on the Neset penninsula at the lake Svardalsvatnet, and the recordings to the south of the Osen inlier. The position of the axial trace is thus also here located quite well. Further to the east the axial trace has been extrapolated in essentially the same direction as established for the rest of the syncline. The axial trace in the Osstrupen syncline is thus a fairly straight line. Also folds in the Hornelen and Kvamshesten Devonian massifs show similar straight axial traces (Bryhni & Lutro 1991a, 1991b), indicating that this is a characteristic feature of the folds in the Devonian massifs.

Due to the plunging fold axis in the syncline, the dip of bedding is slightly shallower in the fold closure (ca. **53°**) than in the limbs (ca. **62°**). The dip of bedding along the axial trace is reflected in the axial trace-profile (**Plate 2**).

Subcylindrical fold

The stereographical girdle formed by all bedding poles from the HDM has been shown to define a π-circle with an orientation of **025/37 NW** (**Fig. 5.51a** and **5.51b**). As much as **90 %** of the poles of the girdle fall within **20°** from the selected average great circle, and the Osstrupen syncline is thus subcylindrical according to the criteria suggested by Ramsay & Huber (1987). Also the fold closures located to the south of Gravanesholten and to the south of Nonsnipa at Osen have pole girdles where **90 %** of the poles fall within **20°** from the selected π-circles (**Fig. 5.52a, 5.52b, 5.53a** and **5.53b**). It is worth noting that the spread is possibly more due to unsystematic variation in the *primary* orientations of bedding from place to place than to later *tectonically* induced systematic variations in the plunge of the fold axis, i.e. to variations produced during the later folding, but precise information on this cannot be obtained due to the large size of the folds, the lack of continous and clearly defined bedding marker horizons, and to the fact that the amount of primary variation cannot be precisely estimated.

The local fold axis of **112/58** obtained from the plateau south of Gravanesholten, and the local axis of **105/50** from the Osen area, both lie within **10°** deviation from the average Osstrupen syncline fold axis of **115/53**. Furthermore, the constant bedding orientations in the limbs suggest that the fold axes elsewhere along the axial trace also do not deviate more than up to about **10°** from the average. This confirms that the orientation of the fold axes is constant along the axial trace.

Concluding remark on the Osstrupen syncline

It has been shown that the Osstrupen syncline has a fold axis dipping around **53°**, and limbs dipping about **50°** when measured in a plane orthogonal to the fold axis. This shows that the fold axis is among the steepest plunging ones, and the fold opening among the closest ones for all the Devonian massifs in Western Norway.

5.5.2.3 DISCUSSION: ***ONE*** SINGLE SYNCLINE VERSUS ***SEVERAL*** LARGE-SCALE FOLDS IN THE HDM

Introduction: In the above description of the Osstrupen syncline (Sect. 5.5.2.2), it has been concluded that the HDM forms *one* single, large syncline. This strongly contradicts Torsvik et al. (1987) (paper enclosed in **Appendix C** of the present thesis), which in their paper concluded that the HDM contains — not one — but *several large-scale folds*; delineated by the form of the Devonian/substrate *contact*. In addition, although not mentioned by Torsvik et al. (1987), a similar conlusion had been suggested by N. H. Kolderup three decades earlier, in the regional geological guide of Kolderup (1960). Since the claimed "several large-scale folds" of Torsvik et al. (1987) is not compatible with the documented presence of the *single* and uniform Osstrupen syncline reported in the present work, the folding-issue will be discussed in the following. The regional geological guide of Kolderup (1960) only briefly mentions the folding of the HDM, and this publication will therefore not be discussed here. The main focus will be on the conclusions in Torsvik et al. (1987), since (1) these are the geologically most wide-ranging, (2) the paper as such deals with the HDM, and (3) the paper is the *most recent* publication focusing solely on the HDM.

As a basis for the discussion it is necessary first to give descriptions of two particular Devonian/substrate contact areas which serve to illustrate the problem and which are referred to by Torsvik et al. (1987) and Kolderup (1960) as the basis for their conclusions; (1) *western areas*, i.e the western margin of the HDM, and (2) *eastern areas*, i.e. the eastern margin of the massif (immediately west of the lake Vassetvatnet) (**Plate 1**). In the descriptions of these areas, the emphasis will be on the large-scale geometry (shape) of the subaerial Devonian/substrate *contact*, on the orientation of the adjacent *bedding* in the HDM, and in particular on the relationship between these two factors. The features described can be seen on the main map (**Plate 1**). At the end of each of the two area-descriptions, the interpretation of these areas in the *present work* is indicated. Subsequently we proceed to the discussion of the conclusions in Torsvik et al. (1987).

Western areas: In the western areas, the marginal Devonian/substrate *contact* defines a simple, trough-like "synclinal" structure. When seen from the sea shore of the westernmost hinge area of the fold, the contact between Devonian rocks and substrate generally climbs steadily N- and S-wards towards higher altitudes, without being off-set or disturbed by faults or palaeotopography to any significant extent (**Plate 1**). A closer look at the Novene area in the west-southwest shows that the subaerial surface trace of the contact there defines a *straight* line all the way from the sea and across the elevated ridge of Novene, indicating that the contact between the Devonian and the substrate must be close to vertical along this distance.

The overall orientation of *bedding* in the west was presented in the previous section (Sect. 5.5.2.2), and forms the western fold closure of the Osstrupen syncline. On the southern side of the axial trace, in the Novene area, the contact dips steeper than the bedding (and this is the only place in the HDM where such a bedding-contact angular relationship has been observed). On the northern side of the axial trace, it is likely that the bedding dips steeper than the contact, and this situation appears to be the normal situation in the HDM.

In conclusion, despite some difference in the angle between bedding and the overall contact surface, both the *contact* and the *bedding* form a simple, large trough structure in the west.

In the present work, this trough structure is interpreted to comply with the presence of a single syncline in the HDM, the Osstrupen syncline.

<u>*Eastern areas*</u>: In the eastern areas, the situation is generally very different, as the *Devonian/substrate contact* does ***not*** form one, single trough structure. Instead, the eastern contact, along its overall N-S directed trend in map view, in detail shows a highly sinuos-curved shape that defines a very complex contact pattern. This can be well illustrated from the steep mountain slopes and valleys to the west of the lake Vassetvatn, i.e. from the northern part of the Svardal area to the Vikafjell area, which belongs to the northern limb of the Osstrupen fold. Along this travers, the contact between the Devonian rocks and the substrate repeatedly climbs and drops in altitude. A difference in altitude of as much as **400 m** has been measured at the southern margin of the Stigen Devonian spur from the lake Vassetvatn and up to the area between Leirvågfjellet and Nonsnova (**Plate 1**). In addition to the altitudal undulation, the contact, as mentioned, also undulates towards east and west in the map plane, around the Devonian spurs.

The orientation of Devonian *bedding* in these eastern parts may be observed along a N-S directed traverse from the summit of Vikanipa to Teigafjell, i.e. a traverse located just to the west of the eastern Devonian/substrate contact. All the way, the *bedding* has a constant strike in the usual ENE-WSW direction and a constant dip towards SSE of the usual magnitude. The strike and dip of bedding thus show no variations along the traverse which could be related to the rise and fall in altitude of the eastern contact, i.e. the changes in the shape of the contact give no corresponding change in the bedding orientation. Furthermore, in large areas, the bedding meets the Devonian/substrate contact at a very high angle, showing that the bedding is *not* generally parallel or subparallel to the contact.

In conclusion, the morphology of the Devonian/substrate *contact* in the eastern part of the syncline (northern limb) is not as simple as in the west, as it does not follow a simple synclinal "fold" morphology. The *bedding* displays a constant strike and dip in the whole area, usually having a large angle towards the Devonian/substract contact.

In the present work the rise and fall of the contact in the east is explained as the result of a combination of palaeo-topography and faulting, but since the present sections merely deals with the geometry of the Osstrupen syncline as such, the detailed discussion of a general structural model for this area is posponed to

Sect. 5.5.8 where the effects of palaeo-topography and faults are treated in relation to the overall structural framework. However, the basis for the rejection of the "several large-scale folds"-model will be discussed in the following.

Discussion: Torsvik et. al. (1987) concluded that the Devonian rocks of Håsteinen and the immediate substrate are folded into *"several large and high amplitude folds"* which over the whole massif *"integrate into a large-scale synclinorium"*. Torsvik et al. (1987) furthermore drew the very general conclusion that *"the form of the folds and the outcrop pattern is defined by the prominent surface of unconformity"*.

However, despite the rigidity of their conclusions, Torsvik et. al. (1987) unfortunately presented no evidence in support of the conclusion that these "several large-scale folds" are actually present. Moreover, the conclusion of Torsvik et. al. (1987), that the form of the claimed folds is defined by the surface of unconformity, was only inadequately substanciated by referring to a figure (scetch map) showing the subparallelism between bedding and the Devonian/substrate contact in the *westernmost area*, (i.e. in the area between the ridge of Novene in the south and the summit of Mannen in the north), an area which on the figure referred to by Torsvik et al. (1987), are shown to contain — not several — but only *one* large fold, the Osstrupen syncline of the present work. Despite the lack of documentation, the conclusion of Torsvik et. al. (1987) (that the HDM contains several large folds delineated by the form of the unconformity), was stated firmly with no reservations, and a discussion of the subject is therefore appropriate.

It should be mentioned that also Kolderup (1960) gave no documentation or evidence to support his conclusion that the HDM contains large-scale folds delineated by the form of the Devonian/substrate contact; a conclusion lying implicit in the claim of Kolderup (1960) that the "arc"-form of the Devonian/substrate contact between the Stigen and Strupeneset Devonian spurs reflects the folding of the HDM.

Although Torsvik et al. (1987) did not give information on the locations of their claimed folds, the fold locations can be easily found from their reasoning: from the above conclusions of Torsvik et al. (1987), and their reference to the western margin of the HDM, it is clear that the authors consider Devonian/substrate contact segments with "syncline"- and "anticline"-shapes to be large-scale folds. Following this fact, the undoubtedly best examples of such "contact-folds" in the whole HDM is the Devonian/substrate contact between and around the Devonian spurs in the eastern part of the HDM, an area where the contact as such has, at several places, an obvious "fold-resembling" rise and fall in altitude. Therefore, although Torsvik et al. (1987) did not explicitly give the geographical locations of their claimed folds, it is obvious that the rise and fall in altitude of the contact around these spurs delineate their claimed "*several large and high amplitude folds*". Also Kolderup (1960) referred to this area when he suggested the presence of several large folds in the HDM.

Contrary to this, as briefly indicated above, the data provided in the present work show that the fold model of Torsvik et al. (1987) is not correct for the areas to the west of lake Vassetvatnet. Because the bedding orientations in each limb of the Osstrupen syncline is constant also in other parts of the HDM, the model of Torsvik et al. (1987) will not apply to other parts of the HDM either. If we, for the sake of illustration, imagin that the rise and fall of the contact in the eastern parts was related to repeated folding, as suggested by Torsvik et al. (1987), the parallelism in the orientation of the Devonian bedding would consequently imply folding of an *isoclinal* type with complete transposition of the layering in repeated folds. Based on detailed work in the area, however, it appears very unlikely that the Devonian spurs and the large variation in the altitude along the contact in the eastern parts of the study area, are caused by such isoclinal folding. The reasons for this are four-fold: (1) the constancy in bedding orientation along a N-S profile through the limb; i.e. no indications of the claimed fold closures, (2) a complete absence of a visually detectable axial planar tectonic fabric, with no tectonically flattened pebbles, etc., which *would* have been present if the folding was so strong, (3) the open continuous fold in the west, which suggests that intense large-scale folding is unlikely in the east (see below), and (4) the generally constant bedding orientations along transects from west to east, i.e. from the open western fold. It may be noted that in the HDM deposits, sedimentary grading is not a useful criteria to exclude or confirm the presence of the postulated opposite limbs in isoclinal folds, since both normal and reverse grading is present in the sedimentary facies in the HDM.

If such an open fold, as has been demonstrated in the west (this work), should be combined with the claimed repeated *isoclinal* folding that was suggested by Torsvik et. al (1987) in the east, it would imply an extreme difference in degree of N-S contraction and structural style between areas separated by only about **5 km** along the axial trace, and between areas that are equally little deformed on mesoscopic scale. None of the other Devonian massifs in western Norway show this kind of extreme strain differences along their axial traces. On the contrary, the other massifs are characterised by a very constant and consistent amount of N-S shortening along the E-W directed axial traces.

Furthermore; the general conclusion of Torsvik et al. (1987) that the form of the folds is defined by the prominent surface of unconformity, and the accompanying reference to the western margin of the HDM, shows that the authors assumed that (i) the Devonian/substrate contact is largely parallel or subparallel to the Devonian bedding throughout the whole massif, i.e. also around the eastern margins of the Devonian spurs in the east, and that (ii) the bedding-contact subparallelism in the western areas is an example proving this. However, as indicated above and seen on **Plate 1**, the data show that the bedding-contact subparallelism in the western part of the Osstrupen fold cannot be applied as an anlogue to the contacts around the Devonian spurs in the east.

Everything considered, the data clearly show that the conclusions drawn by Torsvik et al. (1987) and Kolderup (1960) of several large-scale folds in the HDM, delineated by the Devonian/substrate contact, must be rejected, and it must instead be concluded that the large syncline structure present in the western parts of the HDM continues in the eastern parts, even though the contact in the east has a more complex development due to palaeotopographic relieff during deposition, and minor later faulting (Sect. 5.5.8).

5.5.2.4 SMALLER-SCALE D_1-DEFORMATION FEATURES IN THE HDM

The present section (5.5.2.4) gives a brief review of the status regarding small-scale structures throughout the HDM.

Axial plane cleavage

Axial planar cleavage has been observed in the sandstone facies of the Gravaneset, Vikaholmen and Novene sandstone units in the far west. The three small localities are the only places where axial planar tectonic fabrics have been observed throughout the HDM. In the Gravaneset unit, an axial planar cleavage, and possibly tectonically elongated or flattened pebbles, have been developed in relation to the parasitic folds at Gravaneset. Both the folds and the related axial planar cleavage will be described in detail in the next section (Sect. 5.5.3). In the conglomerates throughout the massif, it is generally not possible to observe any tectonic elongation of the clasts in response to the folding of the HDM into the Osstrupen syncline. The clasts do not show cracking, suggesting that the pressure and temperature was sufficient for ductile deformation.

In Sect. 5.5.2.3, the conclusion by Torsvik et al. (1987) (see **Appendix C**), that the HDM contains, not one but several large-scale folds that are delineated by the form of the Devonian/substrate contact, was rejected on the basis of a comprehensive amount of contradicting data presented in the present thesis. However, in their paper, Torsvik et al. (1987) also concluded that *"in the cores of the larger folds"* [i.e. the rejected ones] throughout the HDM, the compressive orogenic strains were sufficiently penetrative to be marked by some degree of *"preferred orientation of pebble axes"* that were said to be parallel to an *"axial plane cleavage observed in small siltstone lenses"*. Like for the claimed "large folds", no data or locations were given to document the presence of the claimed "pebble fabrics". However, in the similar way as the "*cores of the large folds*" cannot be observed, since the "large folds" are not present (Sect. 5.5.2.3), neither has a preferred orientation of pebble axis in these "cores" (or elsewhere) been observed throughout the HDM. Likewise, none of the observed small sandstone lenses throughout the HDM were seen to contain a cleavage. The "*cleavage in siltstone lenses*" referred to by Torsvik et al. (1987) could be the axial planar cleavage that is present in the marginal sandstone units in the far west, but the paper lacks documentation on this (see Sect. 5.5.3 for a description of the cleavage). However, as mentioned above, this fabric is confined to the sandstone units and related to the local parasitic folding at Gravaneset, and not to the "several large folds" mentioned by Torsvik et al. (1987).

In conclusion, the deformation that has been effecting the HDM has thus not been strong enough to generate neither cleavage nor pebble-flattening / pebble axis-alignment throughout the massif.

Compaction cleavage

Conclusive evidence for the existence of a bedding-parallel compaction cleavage has not been found in the HDM. In the Hornelen Devonian Massif, such a cleavage is present in siltstones and mudstones (authors own observations). Torsvik et al. (1988) have recorded both a conventional tectonic cleavage and a magneto-tectonic "cleavage" from the Hornelen massif, which supports this observation. In the most fine-grained parts of the Gravaneset sandstone unit, a few observations of vague films of microcrystalline white mica might indicate the presence of a compaction cleavage. The observations are, however, not conclusive. Since no convincing evidence for a bedding-parallel compaction cleavage has been found, the folding producing the Osstrupen syncline is considered to be the first phase of deformation (D_1) effecting the HDM.

Parasitic folds

Folds that are parasitic to the large Osstrupen syncline are found only in the Gravaneset sandstone unit in the westernmost part of the Devonian sediments. The folds are developed in the lower part of the deposits; i.e. from the unconformity and about **40 m** up into the Devonian strata as measured normal to bedding. The folds are located about **100 m** to the north of the axial trace of the Osstrupen syncline. A description of the folds will be given in Sect. 5.5.3.1.

It is worth noting that the parasitic folds occur very close to the axial plane of the Osstrupen syncline, and in the lowermost part of the deposit, which is the outermost part of the fold hinge. The parasitic folds also contain an axial planar cleavage (Sect. 5.5.3.1). The hinge area of the Osstrupen syncline has thus experienced layer-parallel contraction.

Pebble-pebble interpenetration

Convincing field evidence of pressure solution and resulting interpenetration of clasts in the conglomerates have not been found. The absence of interpenetration is consistent with the fact that the deformation is generally not strong enough to flatten the pebbles to a degree which is visually detectable. A small degree of pressure solution at clast contacts may be present on the microscopic level (Sect. 5.5.3).

Quartz veins

Quartz veins are virtually absent throughout the HDM. The general lack of quartz veins is in accordance with the lack of extensive pebble-pebble pressure solution and the general low degree of deformation. It appears likely that only very small amounts of quartz solutions were produced from the HDM itself during the deformation/high temperature in Devonian times.

The lack of quartz veins may also indicate that the HDM either did not have fractures that could be filled with vein material during the deformation, or that silica-saturated solutions did not sirculate at this time. If fractures did not develop during the deformation of the HDM, this may indicate that pressure and temperature were relatively high, or that the deformation was slow, so that the deformation occurred in a ductile manner. If hydrothermal solutions did not sirculate, this may mean that joints were not interconnected, or that such solutions did not form.

5.5.2.5 COMPARISON OF STRUCTURAL ELEMENTS OF THE HDM AND THE SUBSTRATE

The present comparison (5.5.2.5) will provide an overall picture of the similarities and differences of the structural elements of the HDM and HC rocks.

The major F_1-fold axis of the HDM has an ESE-directed azimuth and a plunge of **53°** (**Fig. 5.51b**). The minor F_2-fold axes in the substrate, which are cut by the unconformity, also have ESE directed azimuths. The plunges of these F_2-folds are, however, highly variable, although mostly moderate to steep towards the ESE to SE (**Fig. 4.38a**, **4.39a**, **4.40a**, **4.41a**, **4.43a**, **4.45a**). The orientation of the F_1-fold axis in the HDM and the minor F_2-axes in the substrate are thus generally subparallel.

The axial plane in the Osstrupen F_1-syncline strikes WNW-ESE and has a vertical dip (**Fig. 5.54**). The axial planes to the minor F_2-folds in the substrate also have WNW-ESE oriented strikes (**Fig. 4.38a**, **4.39a**, **4.40a**, **4.41a**, **4.43a**). The dips are, however, more variable. In the northwestern, northern, northeastern, west-southwestern and western areas, most axial planes dip steeply to the south (**Fig. 4.38a**, **4.39a**, **4.43a**, **4.44a**). In the eastern and southeastern areas, the planes dip steeply towards both the south and the north, and are thus centered around a vertical dip.(**Fig. 4.40a**, **4.41a**). The data thus show that the Osstrupen F_1-syncline and the substrate F_2-folds have axial planes which also show a general parallelism.

Since the substrate F_2-folds are cut by the sub-Devonian unconformity, the parallelism of the structural elements in the HDM and the substrate indicate a fairly parallel pre- and post-Devonian direction of maximum shortening. The Devonian compression which produced the F_1-fold in the HDM must also have had some effect on the pre-Devonian F_2-folds in the substrate, although these effects have been difficult to identify. These issues are further discussed in Ch. 6.

5.5.2.6 COMPARISON OF THE OSSTRUPEN SYNCLINE (HDM) AND THE GRØNDALEN SYNCLINE (HORNELEN DEVONIAN MASSIF)

All the Devonian massifs of western Norway have been folded into one or more synclines, but the HDM appears to be the strongest folded of all these massifs. This makes the Osstrupen syncline a remarkable structure among the Norwegian Devonian massifs, and thus it is of interest to compare the fold with a syncline that is as similar as possible, in a different massif. In the Hornelen Devonian Massif, the Grøndalen syncline, situated about **10 km** north of the HDM, along the southern margin of the Hornelen massif (**Fig. 5.55**), is suitable for such a comparison. This syncline is the one situated nearest to the HDM, and it also shows several similarities to the Osstrupen syncline, as listed in teh following:

1. The dimensions of the Grøndalen syncline, i.e. the width of the fold, is in a scale of **several kilometres**.
2. The limbs are fairly planar.
3. The northern limb strikes towards the NE (orientation **~ 060/60 SE**), and the southern limb towards the SE (orientation **~128/50 NE**).
4. The dip of bedding in the limbs is up to **50-60°**.
5. The hinge zones are very narrow or absent. (Note that the hinge zone-resembling curved bedding-trace to the east of lake Grøndalsvatnet (**Fig. 5.55**) is *not* a hinge zone, but a result of the intersection of bedding and topographic relief).
6. The fold axis always has an eastward azimuth; although with a shallow to moderate plunge (**azimuth/plunge ~ 087/38**), instead of the moderate to steep plunge of the Osstrupen syncline.
7. The interlimb angle is quite similar (Osstrupen **102°**; Grøndalen slightly variable: **117°**west of lake Kaldevatnet and **97 °** between the lakes of Nipevatna; with a mean value of **107°** (values obtained stereographically).
8. The axial trace forms a fairly straight line.
9. The orientation of the axial trace is essentially E-W
10. The axial plane are dipping steeply to the S, i.e. close to the vertical plane of the Osstrupen syncline.
11. The bedding orientation changes abruptly from one limb to the other.
12. At the southern margin, bedding terminates with a high angle towards the margin, which is defined by a fault.

The great similarity between the Osstrupen syncline of the HDM and the Grøndalen syncline of the Hornelen Devonian Massif may suggest that the style of the deformation was very similar in the two massifs. However, there appears to be one important difference between the massifs: With exception of the western contact, which is an unconformity, the Hornelen massif appears to be situated with a fault contact directly on the subjacent Nordfjord–Sogn Detachment Zone. This differs from the Håsteinen massif, where the entire unit appears to rest with a primary unconformity on the Upper Plate rocks. This is further discussed later. (See Ch. 6).

5.5.3 MINOR D_1-STRUCTURES AT GRAVANESET

5.5.3.1 PARASITIC F_1-FOLDS

General

The massif-wide F_1-folding in the HDM has produced structures of local and minor scale in the Gravaneset area. A set of first order parasitic F_1-fold structures to the Osstrupen syncline, and an accompanying axial plane S_1-cleavage, are developed. No other parasitic folds are present in the HDM. A structural analysis has been carried out, and stereographic representations of the important structural elements are presented.

The presice location of the folds within the larger Gravaneset area can be seen on the 1:1000 map of the westernmost areas of the HDM (**Plate 4**). The exposed parts of the folds are located within an area of about **100 x 100 m**. This area is covered in detail by the 1:200 map (**Plate 5**) which is an enlargement of the framed section on **Plate 4**. All structural data plottetd in the stereograms, and most of the structural elements and features discussed in this chapter, appear on the 1:200 map (**Plate 5)**. The folds are located in a steep escarpment, making access difficult.

The fault crossing the area covered by **Plate 5** has displaced the primary contact between the Devonian strata and the substrate, giving **45 m** of apparent dextral offset in map view. There is no genetic link between the folds and this late fault. The folds are located to the south of the fault, and the data used in the general treatment of the folds, and in the stereograms, are taken from the same area. Data from both sides of the fault are, however, used in the treatment of the axial planar cleavage.

Description of limb orientations

The orientation of the limbs will be briefly described, to draw attention to the — in a Devonian context — spectacular shape of the folds. A conceptual three dimensional sketch of the folds is given on **Fig. 5.56**, indicating the orientation of the limbs. When the fold structure is considered from the north to the south, the orientation of bedding starts with a strike of about **020°** and a corresponding eastward dip of **55–65°** (**Plate 5**). The strike direction of **020°** corresponds to the general strike orientation of the northern limb in this part of the Osstrupen syncline (**Plate 4**). In the *first* fold hinge zone, the strike orientation of bedding gradually rotates clockwise from **020°** to an orientation of **085-095°** with a corresponding southward dip of **60–70°**, producing the first anticline. Across the *second* fold hinge zone, the strike of bedding gradually rotates from **085–095°** and back again to about **015°** with a corresponding eastward dip of **55–65°**, to produce the corresponding syncline of the fold pair. In the field, the folds in the Gravaneset sandstone unit are clearly visible due to the prominent marker horizons defined by the planar bedding in the sandstones as well as by the interlayering of sandstones and conglomerates (**Fig. 5.57a, 5.57b**).

Further towards the south, no outcrops are present for about **50 m**. The 1:200 map (**Plate 5**) therefore does not continue further southwards. The outcrops which are present south of the **50 m** unexposed area, are confined to the road cut. In this road cut, the strike orientation of the Devonian bedding has once again changed, now from **015°** to **075–085°** with a corresponding southward dip of **85°** (**Fig. 5.56**, **Plate 4**). Another anticline must therefore be present between the outcrops in this road cut and the area covered by the 1:200 map (**Fig. 5.56**). This anticline is termed the "unexposed anticline" in the following dicussions.

Only the fold closures of the northernmost anticline and the syncline following to the south are thus exposed (and the degree of exposure is very high). This means that only the folds located on the 1:200 map

(**Plate 5**) have been suitable for a structural analyses, and the following treatment is essentially confined to the area covered by the 1:200 map (**Plate 5**).

Southward from the southern margin of the 1:200 map area, the orientation of the southern limb of the syncline is uncertain due to the **50 m** interval without exposures. It is likely, however, that the limb continues with the same strike orientation of around **015°** towards the southernmost unexposed anticline, i.e. with a strike towards S-SSW. This strike orientation would be consistent with the strike of about **020°** for the north limb of the anticline. Although it is theoretically possibel that the unexposed part of the southern limb segment (of the syncline) may instead have a strike continuing more towards the SSE or even SE, this is not considered likely. Such a limb orientation would imply much stronger folding in the syncline than in the anticline, and the syncline would then become *too* tightly folded to be in harmony with the only vaguely developed cleavage and the lack of pebble flattening in the rock.

From the maps (**Plate 4, Plate 5**), and **Fig. 5.56**, it may be seen that the anticline and the syncline together form a fold-pair which apparently defines an open ***Z-fold*** when viewed towards the east (down-dip). Since the Z-fold is located to the north of the fold-closure/axial trace of the large Osstrupen syncline, it may theoretically be claimed that the fold, in order to have a normal vergence, should have been an ***S-fold***, instead of a Z-fold when viewed E-wards (Pumpelly's rule; Ramsay & Huber 1987). On the other hand, the neighboring fold-pair to the south, consisting of the syncline together with the unexposed anticline, defines a "correct" ***S-fold***. This fact, together with the general low degree of folding in the HDM, suggest that the apparent "wrong vergence" indicated by the Z-fold, is merely a result of local *irregular* parasitic folding near the fold closure of the open Osstrupen syncline, and *not* a feature indicating an important deviation from "normally vergenced" parasitic folding. As we shall see below, also the *axial plane* orientations at Gravaneset deviate slightly from that of the Osstrupen syncline, i.e. further illustrating and confirming this minor irregularity of the Gravaneset folds.

Fold axes

The fold axes have been found from π-plots of poles to bedding. The anticline has 36 poles defining a girdle (**Fig. 5.58a**), and the best fitting great circle on the contoured stereogram gives a *mean fold axis oriented **125/55*** (**Fig. 5.58b**). Also the syncline has limb poles defining a girdle (**Fig. 5.59a**), and the best fitting great circle here yields a *mean fold axis oriented **129/59*** (**Fig. 5.59b**).

Mutually, the two fold axes of **125/55** and **129/59** in the folds at Gravaneset are very close in orientation, and also fairly close to the fold axis of **115/53** for the *Osstrupen syncline*.

Interlimb angles

The interlimb angles given below are found from the stereographic techniques shown in **Fig. 5.60** and **5.61**. The bedding/limb orientations used to represent the average maximum closure of the Gravaneset folds are taken from the best fitting π-great circles to the bedding poles (**Fig. 5.58b, 5.59b, 5.60, 5.61**). Since the bedding-strike changes gradually along the limbs, the representative limb orientations yielding maximum fold closure are taken from the marginal parts of the pole populations along the great circles. For the anticline these limb orientations are **090/68 S** and **010/58 ESE** (**Fig. 5.60**), and for the syncline they are **090/69 S** and **018/60 ESE** (**Fig. 5.61**). These limb orientations produce an *interlimb angle of* ***110°*** *for the anticline and* ***115°*** *for the syncline* (**Fig. 5.60, 5.61**). The folds thus have very similar opening angles. The folds have interlimb angles in the interval **70-120** has been observed is in the sandstone facies of the Gravaneset, Vikaholmen and Novene sandstone units in the far west. **°**, and are thus classified as "open folds" according to the criteria suggested by e.g. Ramsay & Huber (1987).

The interlimb angle for the *Osstrupen syncline* was estimated to **102 °** (Sect. 5.5.2.2). The interlimb angles of the folds at Gravaneset (**110** and **115°**) are deviating only **8°** and **13°** from this value, i.e. the interlimb angles are thus quite similar to that of the *Osstrupen syncline*.

Axial planes

The orientation of the axial planes in the folds at Gravaneset can be estimated using the same stereographic procedures as for the axial planes of the Osstrupen syncline (Sect. 5.5.2.3). Also in the Gravaneset folds the axial plane is taken to correspond to the bisecting plane of the interlimb angle. This leads to the results that the *axial plane for the anticline is oriented* ***136/82 NE*** (**Fig. 5.60**), and the *corresponding plane for the syncline is oriented* ***140/83 NE*** (**Fig. 5.61**). The two axial planes of the folds at Gravaneset thus have parallel orientations, as have the two fold axes. The fold axes with orientations **125/55** (anticline) and **129/59** (syncline) lie within these axial planes (**Fig. 5.60, 5.61**).

The axial plane of the *Osstrupen syncline* is oriented with a strike/dip of **115/90** (Sect. 5.5.2.2). The *dips* of the Gravaneset axial planes thus differs only **7°** and **8°** from the axial plane of the Osstrupen syncline, a difference so small as to be negligible. The *strikes* of the Gravaneset axial planes differs **21°** and **25°** from the Osstrupen syncline, and this difference is just large enough to be significant.

Discussion on axial plane orientation: The slight difference in orientation between the Gravaneset and Osstrupen axial planes deserves a discussion: The axial planes of the folds at Gravaneset have been constructed based on the assumption that they correspond to the planes bisecting the interlimb angles ("bisector planes"). Theoretically, this geometric construction of the axial planes may be incorrect, since the possibility exists that the axial planes may in fact deviate somewhat from the bisecting plane, so that the orientations of the Gravaneset axial planes may be closer to the Osstrupen axial plane than they appear to be from the stereographic plot. However, for the present folds, the assumption that the bisecting planes correspond to the axial planes must nevertheless be correct, and the reasons for this are as follows: to produce a significant separation of the axial plane from the bisector plane of a fold, the two limbs involved must be subjected to very uneven limb distortion, i.e. one of the limbs must be significantly more thinned/attenuated than the other (Turner & Weiss 1963; Davis & Reynolds 1996). At Gravaneset, however, the very moderat folding has only given a gentle bending of the layers, and no visible limb-attenuation. It is therefore highly unlikely that one of the limbs should be significantly more attenuated than the other, and therefore unlikely that the axial planes should deviate much from the bisector planes. Since the folding at Gravaneset has only given carefully bent limbs with no signs of attenuation of any particular limbs, the geometrically constructed axial planes must be very near the real axial planes. The deviation of the Gravaneset axial planes from the Osstrupen axial planes is small and just barely significant.

The validity of the above constructions, yielding the orientation of the axial planes, depends on the two conditions that (1) the measurements used to represent the limb orientations must be representative, consistent, and of high quality, and (2) *one* of the limbs must not have been significantly more attenuated than the other. Both these conditions are met in the present case. The low degree of deformation, i.e. the fact that the folds are *open* folds instead of *tight* folds, is therefore not in itself a factor making the construction subject to large errors, as long as the other two conditions are met.

The slight deviation of the Gravaneset axial planes from the Osstrupen axial planes is probaly the result of local stress and competence conditions which have led to slightly irregular parasitic folding near the fold closure of the Osstrupen syncline. The relativly limited amount of deformation across the Osstrupen syncline makes such irregular folding likely and reasonable, as the deformation on the HDM was *not* strong enough to rotate the parasitic axial plane-strikes, fold axes-trends, etc., into parallelism with those of the large fold structure, thus resulting in a slight "fanning" of these minor structural elements. The apparently "wrong vergence" of the fold mentioned above should not be considered as a *result* of the deviation of the Gravaneset axial planes from the Osstrupen axial planes. Rather, it is more correct to say that *both* the "wrong vergence" and the deviation of the Gravaneset axial planes, are due to slightly *irregular* folding at the closure of the Osstrupen syncline.

Axial traces

We have seen that the axial planes were found by stereographic construction based on the assumption that the axial planes correspond to the planes bisecting the interlimb angles. The axial *trace* is defined as the position of the axial plane (surface) on the ground surface (Ramsay & Huber 1987, Davis & Reynolds 1996). The orientation of the axial traces of the anticline and the syncline at Gravaneset is indicated on the 1:200 map (**Plate 5**). The traces are here gently curved due to the interaction with the topography, and have strike directions ***around*** ***136°*** (anticline) and ***140°*** (syncline).

When, as in the present case, the fold axis has a plunge and the axial plane is inclined, the axial trace and the axial trend (hinge line azimuth) will not be parallel (Ramsay & Huber 1987). The hinge line trace will be parallel to the fold axes trace, and will thus be trending **125°** and **129°** for the folds at Gravaneset, i.e. diverging only **11°** from the axial traces. The small divergences between the axial trace and the axial trend is due to the steep dip of the axial planes (**82°** for the anticline, and **83°** for the syncline).

The axial trace of the *Osstrupen syncline* has a trend around **115°** as inferred from the vertical axial plane with its strike direction around **115°**. This trend is **21°** from the trend of the axial trace of the Gravaneset anticline, and **25°** from the syncline, implying that the trend of the axial traces of the Gravaneset folds is fairly close to that of the *Osstrupen syncline*.

Wavelength / amplitude

The wavelength of the folds is minimum **40 m**, and possibly as much as **100 m**. The amplitude is in the order of **20 m** (**Plate 5**).

Cylindrical fold

The poles to the bedding defining the limbs in the folds have been shown to fall along a girdle (**Fig. 5.58a, 5.59a**). More than **90 %** of the poles in these plots are situated within a **10°** deviation from the best fitting great circles (**Fig. 5.58b, 5.59b**), and in the scale considered, the folds are therefore cylindrical within the area mapped, according to the criteria suggested by Ramsay & Huber (1987).

Congruous fold

The fold axes to the anticline and syncline at Gravaneset have orientations **125/55** and **129/59**, and, hence, are mutually congruous. Although the plunges of the parasitic folds are practically the same as for the main axis of **115/53** for the *Osstrupen syncline*, the azimuth of the two parasitic fold axes deviate **10-15°** from the main axis. However, since the *Osstrupen syncline* has been classified as a subcylindrical fold (Sect. 5.5.2), and the difference in azimuth is still fairly small, a congruous relationship must be said to exist between the parasitic folds and the Osstrupen syncline.

5.5.3.2 AXIAL PLANAR S_1-CLEAVAGE

General

The present section deals with the axial planar S_1-cleavage in the parasitic F_1-folds in the Gravaneset sandstone unit. Axial planar-type cleavage is only present in three small areas in the HDM, all being located within the sandstone units in the westermost areas, notably in the (1) Gravaneset, (2) Vikaholmen, and (3) Novene sandstone units. Of these three, the folded sandstones at Gravaneset is clearly the dominating cleavage-area, and focus will therefore be on this area. The present section describes the cleavage at *Gravaneset* and presents stereographic plots. As we shall see, the cleavage at Gravaneset is actually not perfectly axial planar in orientation, but since the deviation from such an orientation is relatively small, the cleavage is nevertheless termed an axial plane cleavage. The cleavage at the smaller *Vikaholmen* is clearly visible in the field, whereas the cleavage is only poorly developed at *Novene*, and these two areas will here only be mentioned for comparison of cleavage orientations.

Torsvik et al. (1987) have investigated tectono-*magnetic* fabrics in the Håsteinen area, with particular focus on the Gravaneset sandstones. Since these fabrics are closely connected to the *tectonic* S_1-

cleavage described in the present thesis, they are of interest for the present work. The tectono-magnetic fabrics are discussed separately in Sect. 5.5.4.

The fact that an axial planar cleavage is only developed in the sandstone unit, is probably due to a lower competence in the sandstones compared to the surrounding coarse conglomerates. The alternative possiblility, that a somewhat higher compressional stress acted in this part of the Devonian massif, leading to more intense deformation in the area, is speculative and difficult to assess. The large open syncline in the west suggests an overall *even* distribution of strain.

Description

Mesoscopically, the axial planar cleavage is usually weakly to moderately developed. The orientation of the cleavage can, nevertheless, usually be easily measured on the outcrop. In the field, the cleavage normally has the appearance of a vague to moderate slaty cleavage and thus appears to be a "continuous cleavage" according to the classification of Davis & Reynolds (1996). The cleavage is strongest developed in the most fine-grained sandstones. Occasionally it shows a tendency to grade into a fabric approaching a spaced cleavage with a "disjunctive" (discontinous) appearance, as defined by Davis & Reynolds (1996) (**Fig. 5.62**). The spacing generally varies from **0–3 cm** and tends to decrease with decreasing grain-size in the sandstones. The sandstones with the smallest grain-size generally consist of sand grains that are medium to coarse. The deformation has generally been too limited – and the grains generally too coarse – to produce a real slaty cleavage in the rock when assessed on the outcrop.

The Gravaneset sandstone unit consists of alternating layers of sandstones and pebbly/cobbly conglomerates. The cleavage can only be observed in the sandy portions of the unit (**Fig. 5.62**), but although no general tectonic fabric is present in the conglomerates, some clasts appear to be oriented parallel to the cleavage (see below).

At the *microscopic* level, the tectonic fabric is quite prominent and penetrative, considering that the rock has been subjected to such moderate bulk deformation (**Fig. 5.63a**). However, before describing the cleavage as such, it should be mentioned that the relatively low degree of deformation in the rock normally allows recognition of remnants of the sedimentary texture in the cleaved sandstones. Grains with clastic outlines (**Fig. 5.63b**) can still be discerned among the more recrystallised ones. The texture may therefore be termed blastopsammitic, as defined by Spry (1969).

The mineralogy of the sandstone will also be briefly referred before proceeding to the cleavage. It is here convenient to separate the minerals of the sandstone into the two fractions "grains" and "matrix", where, at least, the larger grains can be identified as being of clastic origin. Of the clastic grains, quartz is strongly dominating. In addition, grains of white mica, epidot, feldspar, sphene, zoisite and opaque minerals are abundant. The matrix is typically very fine-grained, consisting mainly of sericite and opaque minerals, but also of chlorite grains. Minerals having grown during the D_1-tectonometamorphic event is further discussed under "M_1-metamorphism" (Sect. 5.5.6).

Although remnants of a sedimentary texture can be discerned, the cleavage is a dominant feature of the rock. As will be illustrated below, all grain sizes, from the larger "clastic" grains, to the minute metamorphic sericite grains in the matrix, contribute to the formation of the cleavage.

The ***grains*** play a vital role in the cleavage formation, as the cleavage is predominantly defined by parallel orientation of the longest dimension of platy grains/minerals (**Fig. 5.63a**). Grains of quartz and white mica are particularly important fabric-generating elements, but also epidote, sphene and opaque minerals contribute significantly (**Fig. 5.63a**).

The clearest expression of the cleavage can be found in the areas where the cleavage is oriented at high angle to the bedding (**Fig. 5.63a**), since it can be excluded in such areas that the preferred dimensional orientation of the platy grains is a primary sedimentary feature or a primary sedimentary feature that has been only slightly tectonically enhanced. The parallel orientation of the grains creates a grain-shape fabric or mineral fabric of S-type, as defined by, for instance Fry (1984) (**Fig. 5.63a**).

Since the grains showing preferred dimensional orientation frequently have preserved remnants of their clastic outlines (**Fig. 5.63b**), it is possible, following Borradaile et al. (1982) and Davis & Reynolds (1996), that grain sliding and rigid body rotation have been responsible for the alignment of the grains. These processes are common in the earlier stages of the formation of slaty cleavage (Davis & Reynolds 1996). On the other hand, Davis & Reynolds (1996) states that most workers now believe that cleavage is largely created by pressure solution (and to a less degree grain rotation), and it is thus likely that pressure solution has taken place.

Nearly all quartz grains have suffered intracrystalline deformation resulting in undulose extinction. Although monocrystalline quartz grains are still widely present, a large portion of the grains have become polycrystalline due to stress-induced polygonisation with formation of subgrains, a process described by for instance Spry (1969), Bell & Etheridge (1973) and Borradaile (1982). It should be noted that the term "subgrain" is here used in a general sense, meaning "small strain-induced recovery-grains", and *not* in the strict sense of Bell & Etheridge (1973) for grains separated from the host with a less than **7°** boundary. The smallest clastic quartz grains have occasionally been almost completely replaced by minute subgrains (**Fig. 5.64**), yielding a matrix-like appearance. In the more typical larger and prominent grains, subgrains have formed predominantly along the rims. The polygonisation often occurs in contact zones between the grains (**Fig. 5.64**) where the strain has been high enough to generate the strain-removing recovery-process, producing the smaller strain-free subgrains. Where polygonisation has occurred, relict cores of quartz grains tend to be somewhat elongated parallel to the cleavage. Preferred dimensional orientation of the quartz subgrains themselves also contributes to the formation of the fabric (**Fig. 5.64**), as has also been described by Borradaile et al. (1982).

The polycrystallinity/subgrains produced by the *Devonian* deformation can in most instances easily be separated from polycrystallinity inherited from the *source area*, regardless of whether the source area polycrystallinity is in the form of real subgrains, or polycrystalline rock fragments. The Devonian polygonisation has produced very characteristic small subgrains that for larger detrital grains are located essentially along the rims of the grains (**Fig. 5.64**). Detrital grains with a polycrystallinity inherited from the source area have subgrains/newgrains that are usually much larger and typically filling the entire grain (**Fig. 5.65**).

The *matrix* also contributes significantly to the cleavage. The aggregates of sericite and minor chlorite frequently have elongated forms, and the longest dimensions have a preferred orientation parallel to the cleavage. This creates an aggregate-shape fabric of S-type, as defined for example by Fry (1984) (**Fig. 5.63a**). Also within the mica aggregates, individual minute mica grains in the matrix aggregates have a preferred orientation of longest dimension (**Fig. 5.66**). In addition, the matrix occasionally contains sericitic films which, along grain-matrix contacts, have grown parallel to the cleavage (**Fig. 5.65**).

Cleavage-parallel films of white mica have frequently grown along the contacts between grains (**Fig. 5.64, 5.65**). The growth is probably enhanced by pressure solution of quartz due to the higher pressures along the grain contacts, a process described, for instance, by Spry (1969) and Davis & Reynolds (1996).

In strain shadows between quartz grains, larger white mica grains have grown with longest dimensions parallel to S_1 (**Fig. 5.67**). Occasionally, fibrous beardlike quartz oriented in the cleavage plane (c.f. Davis & Reynolds 1996) is also present in the strain shadows (**Fig. 5.65**).

Larger individual mineral grains of white mica have occasionally grown with preferred dimensional orientation outside strain shadows (**Fig. 5.68**).

Detrital clastic grains of white mica and chlorite have occasionally been sharply bent or almost “broken”, so that one leg stands with a high angle to the cleavage, and the other is practically parallel to the cleavage (**Fig. 5.69**). This bending was produced by the deformation, when the grains did not succeed in rotating into the cleavage.

The cleavage-forming processes

By comparison with Borradaile et al. (1982) and Davis & Reynolds (1996), the cleavage-forming processes in the sandstones of the Gravaneset sandstone unit can be summarised as: (1) rigid body rotation, (2) intracrystalline deformation producing recrystallisation, (3) pressure solution, and (4) syntectonic mineral growth.

Although a prominent cleavage is clearly present in the rock, the deformation has not been strong enough to produce the domainal texture charactristic of a fully developed slaty cleavage (c.f. Borradaile et al.1982; Davis & Reynolds 1996). The cleavage in the Gravaneset sandstone unit may therefore be termed an early-stage slaty cleavage. The cleavage is oriented subparallel to the XY-plane of the local strain ellipsoid.

Orientation of the cleavage

From the 1:200 map (**Plate 5**) it can be seen that the cleavage strikes WNW-ESE and dips towards the S. The poles to the **35** cleavage measurements have been plotted stereographically, and a very good cluster is obtained (**Fig. 5.70a**). *The average axial planar-type cleavage has an orientation of **112/65 SSW*** (**Fig. 5.70b**). As much as **97 %** of the measurements lie within a deviation of **10°** in strike and dip from the mean value (**Fig. 5.70a**). The single measurement plotting outside the cluster is from a small locality **30 m** east of the southeastern (i.e. upper) fault-plane/unconformity intersection shown on the 1:200 map (**Plate 5**). The deviating area is negligible in size compared to the rest of the cleaved area. The measurement is considered as a local irregularity with no importantance in the overall picture.

At *Vikaholmen*, the cleavage is oriented around **142/85 SW**, i.e. deviating **30°** in strike and **20°** in dip from the mean value at Gravaneset. At *Novene*, the cleavage is oriented around **126/82 SSW**, i.e. deviating **14°** in strike and **17°** in dip from Gravaneset. Although these deviations are relatively small, they show that local variations occur in the orientation of the cleavage in the three western sandstone units.

Cleavage not completely parallel to the axial planes

The axial planes to the anticline and the syncline of the parasitic folds at Gravaneset have been determined by construction as **136/82 NE** and **140/83 NE**, respectively (**Fig. 5.60**, **5.61**) (planes also shown in Fig **5.70b**), whereas the cleavage has an average orientation of **112/65 SSW** (**Fig. 5.70b**). This means that the axial planes and the corresponding cleavage plane do not completely coincide. The difference in *strike* between the axial planes and the average cleavage is **24°** (anticline) and **28°** (syncline), and the corresponding difference in *dip* is **33°** and **32°**. The interplane angle (profile plane) between the cleavage plane and the axial plane is **41°** (anticline) and **42°** (syncline) (**Fig. 5.70b**). The axial planes dip steeply towards NE, and the axial planar cleavage steeply towards SSW. The following consideration can be made to evaluate whether this difference is scientifically significant, or less than the error bars.

Discussion: The orientation of the axial *planes* have error bars in the order of maximum **+/- 10°**, since the planes appear in cylindrical folds (c.f. Ramsay & Huber 1987). Also the *cleavage* has been shown to have error bars in the order of **+/- 10°** (**Fig. 5.70a**). For both the axial planes and the cleavage, the values selected to represent the *average* estimated orientations will be closest to the "true" values. This appears to indicate that the difference in orientation between the axial planes and the corresponding cleavage plane is scientifically significant, although relatively small. These considerations are based on the condition that the axial planes really correspond to, or are fairly near, the bisector planes.

Two alternatives exist when we shall explain the deviation between the axial planes and the cleavage plane: (1) the deviation between the axial plane and the cleavage plane is either real, or (2) the constructed axial plane is wrong, and the actual axial plane is in reality much closer to the cleavage plane than it appears to be from the stereograms.

The fact that the cleavage poles are so strongly concentrated in *one* population around the average pole, indicates that competence-related effects like cleavage fanning *within* the folds, etc. (c.f. Ramsay & Huber 1987) cannot explain the divergence.

Alternative (2) above must be rejected for the following reasons: In the description of the Gravaneset folds (Sect. 5.5.3.1, headline: "axial planes"), it was concluded that the Gravaneset axial planes essentially correspond to the planes bisecting the interlimb angles of the folds, i.e. to the bisector planes. It was explained that a significant divergence between the axial plane and the bisector plane would require very uneven limb distortion, where one of the limbs would have to be significantly more thinned/attenuated than the other (Turner & Weiss 1963, Davis & Reynolds 1996). It was further argued that for Gravaneset, the deformation has been so limited that the degree of uneven limb distortion necessary to produce a significant separation of the bisector plane from the axial plane (c.f. Davis & Reynolds 1996) very likely has not occurred, i.e. the geometric construction must be very near the actual axial plane. Although the constructed axial planes are essentially correct, it is, off cource, possible that the axial planes in fact deviate *slightly* from the bisecting planes so that the orientations of the actual axial planes are *a bit closer* to the cleavage than they appear to be from the stereographic plot.

Alternative (1) is the most likely explanation, i.e. that the difference in orientation is real, and that presently unresolvable local competence and stress configurations operating during the folding, have produced the difference. If the folding in the Gravaneset area had continued for a longer time, it is likely that the fold limbs would have rotated into a position where the axial plane and the cleavage plane would have been more parallel, and possibly more parallel to the axial plane of the Osstrupen syncline.

The cleavage at *Vikaholmen* with orientation **142/85 SW** is located about **400 m** southwest of the exposed folds at Gravaneset. This cleavage has a strike which is parallel to the constructed axial planes for the Gravaneset folds, a feature which may be taken to indicate that axial planes and cleavages *generally* have slightly varying orientations in the western fold closure of the Osstrupen syncline, indicating that the difference between the axial plane and the cleavage at Gravaneset is an unproblematic feature.

It may be assumed that the Gravaneset cleavage has formed with an orientation close to the XY-plane of the local strain ellipsoid. If so, the axial planes do not correspond completely to the XY-plane of the strain ellipsoid present during the formation of the cleavage and folds. Despite the small difference in orientation, the cleavage is definitely of axial planar type, i.e. the cleavage was formed by the same processes as cleavages which are more *perfectly* axial planar between the limbs of a fold.

Possible rotation of pebbles into the cleavage

Conglomeratic horizons are present within the sandstones folded by the parasitic folds at Gravaneset. In a few of these conglomerates — where bedding is oriented with a high angle to the cleavage — pebbles appear to have their long axes oriented parallel to the cleavage (**Fig. 5.71**). The long axes then make an angle of about **40°** with the bedding. The locality where this is seen is situated about **30 m** east of the southeastern (i.e. upper) fault-plane/unconformity intersection on the 1:200 map (**Plate 5**) — a small site where the cleavage has a more NW-SE strike instead of the normal WNW-ESE strike, i.e. making the bedding/cleavage angle smaller. The clast/cleavage parallelism might indicate that the pebbles have been rotated into the cleavage. It should be noted, however, that in river deposits, primary depositional imbrication-angles of clasts are usually in the range of **10-30°** (Blatt et al. 1980), and in coarse alluvial deposits the angles may reach **40°** (Collinson & Thompson 1989). The general deformation of the folded area at Gravaneset is not very high, and the angle of ca. **40°** between the clast axes and bedding-strike is not clearly larger than the referred common primary imbrication angles in river deposits and coarse alluvial deposits. Therefore, the possibility exists that primary (depositional) imbrication of pebbles could, by coincidence, have received the same orientation as the cleavage. It is also possible that primary imbricated clasts have been subjected to only slight tectonic rotation. The clasts do not appear to have been tectonically flattened.

The interpretation of the clast/bedding angle, as due to primary sedimentary imbrication, is also supported by the fact that the locality is situated at the vicinity of the cleaved area where the cleavage is beginning to die out. The exposures do not allow three dimensional control of the orientation of the actual longest pebble axes to be established, and it has therefore not been possible to find out whether the longest pebble axes define a lineation.

Miscellaneous

Despite the presence of cleavage in the folded Gravaneset sandstone unit, convincing field evidence of pressure solution-controlled, clast-clast interpenetration has not been found in the conglomeratic layers in the unit (although weak signs of pressure solution seems to appear between grains in sandstones). As for the rest of the Devonian massif, the absence of clast-clast interpenetration is also here consistent with the fact that the deformation is generally not strong enough to flatten the pebbles to a degree which is visually detectable. On the microscopic level, a small degree of pressure solution may be present at clast contacts.

Likewise, quartz-veis are virtually absent in the folds at Gravaneset and the surrounding areas, which is again consistent with the lack of extensive pressure solution and the general low degree of deformation.

5.5.3.3 COMPARISON OF STRUCTURAL ELEMENTS IN THE DEVONIAN AND THE SUBSTRATE AT GRAVANESET

Although the axial *planes* and axial planar *cleavage* in the Gravaneset sandstone unit have somewhat different orientations, (Sect. 5.5.3.2), they will be considered as subparallel in the present context. At Gravaneset, the axial planar *cleavage* in the Gravaneset sandstone unit is oriented with a WNW-ESE directed strike and a steep dip (**Fig. 5.70a, 5.70b**). This cleavage is parallel to the *axial planes* of the pre-Devonian F_2-folds in the substrate (**Fig. 4.44a**). In addition, the *fold axes* in the Devonian rocks have WNW-ESE directed azimuths (**Fig. 5.58b, 5.59b**), and are parallel to the azimuths of the *fold axes* of the substrate (**Fig. 4.44a**). The dip of the fold axis is moderate to steep towards the ESE for the Devonian rocks (**Fig. 5.58b, 5.59b)**, but higly variable for the substrate rocks (**Fig. 4.44a**).

Also the *Osstrupen syncline*, with its WNW-ESE striking and subvertical axial plane, and fold axis with a WNW-ESE azimuth and moderate to steep plunge, is parallel to F_2-folds in the substrate. This means that the pre-Devonian strain producing the F_2-folds was coaxial with the post-Devonian strain producing the Osstrupen syncline and the parasistic folds at Gravaneset.

The deformation postdating the deposition of the Devonian sediments must have had some effects on the already existing pre-Devonian structures. However, such effects have been very difficult to identify, and this subject is further discussed in Ch. 6.

5.5.4 TECTONO-MAGNETIC FABRICS

Tectono-magnetic investigations have been carried out in the western part of the HDM and substrate by Torsvik et al. (1987) (**Appendix C**). A large portion of the data have been collected from the Gravaneset area, and particularly from the cleaved Devonian sandstones related to the parasitic folds in the Gravaneset sandstone unit. The investigation of Torsvik et al. (1987) has included both *magnetic fabric* studies and *palaeomagnetic* studies. In the present account, the magnetic fabric studies will be brifly summarised and discussed, since they relate directly to the parasitic folds at Gravaneset and the corresponding axial plane cleavage, which have been documented and analysed in the present thesis (Sect. 5.5.3.2).

Method: The magnetic fabric studies are based on a technique which identifies the orientation and shape of the magnetic susceptibility ellipsoid, which is related to the tectonic strain ellipsoid. When a deformed sample, oriented from the field, is magnetised in different directions in the laboratory, the magnitude of the magnetisation that the rock takes up will vary with directions within the sample, enabling us to define a maximum (K_{max}), an intermediate (K_{int}), and a minimum (K_{min}) magnetisation direction. These directions correspond to the X-, Y- and Z-directions of the strain ellipsoid. Although the *shape* and orientation of of the grains/minerals receiving magnetisation was originally a sedimentary feature, the shape and orientation have later, to a certain degree, been controlled by the tectonic deformation. This leads to a maximum magnetisation (K_{max}) in the direction of maximum elongation, i.e. in the X-direction of the strain ellipsoid. K_{max} is also termed the *magnetic lineation*. Furthermore, the tectono-magnetic K_{max}-K_{int} plane in the susceptibility ellipsoid thus corresponds to the structural XY-plane (or axial plane) in the strain ellipsoid, also called the magnetic foliation.

Results: The investigation by Torsvik et al. (1987) has yielded K_{min}-vectors of fairly constant and consistent orientations. These vectors are, when plotted in a stereogram, poles to the tectonomagnetic K_{max}-K_{int} planes, or the magnetic foliation. The K_{min} "poles" plot in the NNE and SSW part of the stereogram, and the K_{max}-K_{int} planes are oriented with a WNW-ESE strike and subvertical dips (**Fig. 5.72**) (Torsvik et. al.1987). When these data are compared with the poles to the axial *planes* (**Fig. 5.60, 5.61**) and the axial planar S_1-*cleavage* (**Fig. 5.70a, 5.70b**) reported by the present author, it is seen that the range of pole orientation to the magnetic foliation encompasses both the *axial plane poles* and the axial planar *S_1-cleavage poles.* This shows that the orientation of the magnetic foliation presented by Torsvik et. al. (1987) fits the structural data presented in the present thesis. Torsvik et al. (1987) also presented a few single measurements of the cleavage, and these are compatible to the more comprehensive data-set in the present thesis.

As mentioned above, the magnetic lineation K_{max} in the magnetic strain ellipsoid corresponds to the X-direction of the strain ellipsoid. In the magneto-tectonic strain ellipsoid, the average K_{max} is oriented with an azimuth towards the ESE and a plunge of **50–60°** (**Fig. 5.72**). When the K_{max} data are compared with the orientation of the fold axis of the Gravaneset parasitic folds (e.g. **Fig. 5.60, 5.61**), reported in the present thesis, it is observed that the range of K_{max} and the fold axes overlap. This suggests that the fold axes are parallel to the X-direction of the local strain ellipsoid. Torsvik et al. (1987) also presented a cleavage-bedding intersection axis (fold axis) based on only a few measurements. This fold axis was shown to correspond reasonably well to K_{max}, and this relationship is confirmed by the more comprehensive data-set in the present work.

Torsvik et al. (1987) plotted the K_{max}/K_{int} and K_{int}/K_{min} values from **4 sites** in the lower- and western-most part of the Gravaneset sandstone unit, in a Flinn diagram. The apparent shapes of the magnetic strain ellipsoids in the folds at Gravaneset are mainly prolate (**Fig. 5.72**) (Torsvik et. al. 1987).

The degree of tectono-magnetic anisotropy (An) in a rock is given by the expression

$$\% \text{ An} = (K_{max} / K_{min} - 1)\ 100$$

(Torsvik et al.1987). In the Devonian sediments at Gravaneset and the related folds, the anisotropy was found to be typically **5-6 %** (Torsvik et al. 1987).

With this estimate given, the formula allows calculation of the K_{max} / K_{min} ratio. We may set

$$K_{min} = 1$$

to find the magnitude of K_{max} relative to K_{min}. Inserting this number into the formula, together with, for example, the 5% anisotropy, gives

$$5 = (K_{max} - 1)\ 100$$

$$K_{max} = 1.05$$

Consequently, using the anisotropy of 6% would give

$$K_{max} = 1.06$$

This implies that the K_{max}/K_{min} ratio for the magnetic ellipsoid lies between **1.05 : 1** and **1.06 : 1.** Since these ratioes may be transferred to the X:Z of the strain ellipsoid, the ratioes can be considered as a minimum estimate of the contractional strain in the folds at Gravaneset.

A laboratory fold test has been applied to the folded Devonian sediments, and the test was interpreted as negative (Torsvik et al. 1987). This means that most of the folding (in the field) occurred before the magnetism was acquired, i.e. whilst the temperature was high enough to allow the magnetic vectors to reorientate to keep parallelism with the “earth magnetic field”, during the folding of the rocks. For the sandstones at Gravaneset, 70 % of the folding had been acquired before the temperature dropped to a level that caused the magnetic vectors to “freeze” in the rock (Torsvik et al 1987, their Fig. 11e). At this stage, the vectors had parallel directions, regardless of position in the folded rock. After the “freezing” of magnetisation, the last 30% of the folding occurred, leading to a spread of vectors in the affected areas. The authors also made a fold test on data from the Gravaneset sandstones in combination with data from conglomerates at Straumsnes (Torsvik at al. 1987, their Fig. 11f). During this fold test, a reduction in spread of magnetic vectors was not obtained. Instead, an increased spread was seen after only **10 %** unfolding, and this development continued throughout the fold test. Applied to the field situation, this means that practically all Devonian folding of the rocks had been accomplished before the magnetisation was "frosen" into the rock, i.e. that the temperature was high enough to keep the magnetic vectors parallel regardless of rotation (folding) of the rocks. In conclusion, the Gravaneset magnetisation (as well as the Håsteinen magnetisation in general) has thus been imposed *either late syn-tectonically or post-tectonically*, whilst the temperature was still higher than the temperature that led to "blocking" of the magnetisation (Torsvik et al. 1987). The issue, however, of estimating the *actual* temperature that led to "freezing" of the magnetisation in Håsteinen is very difficult, since remagnetisation can occur well *below* the defined blocking temperature (T_B) for a magnetic mineral, as a result of chemical processes controlled by for example hydrothermal fluids. Haematite, that was reported to be the most important remanence carrier in the Håsteinen sediments (Torsvik et al. 1987), has a blocking temperature (T_B) of **~ 650 °C**, whilst magnetite, that is less important in the samples, has a T_B of **~ 550 °C** (Fowler 1990). It is obvious that the maximum temperature in Håsteinen has been much lower than this, but the actual temperature cannot be revealed by palaeomatnetic methods.

The study by Torsvik et al. (1987) also included the magnetic foliation in the *substrate* at Gravaneset. Here, the magnetic K_{max}-K_{int} foliation planes were oriented with NW-SE directed strikes and steep dips, and the poles (K_{min}) were thus very close to the poles to the axial *planes* of the F_2-folds of the Høydalsfjorden Complex (**Fig. 4.44a)**. The magnetic foliation in the substrate was thus also parallel to the magnetic foliation in the overlying Devonian rocks. This parallelism of magnetic foliations was interpreted by Torsvik et al. (1987) to be the result of either an exclusively "Devonian" magnetic fabric in all rocks in the Håsteinen area, or alternatively the superposition of a "Devonian" WNW-ESE fabric on an older but parallel structural grain in the substrate, implying two coaxial phases of N-S shortening.

In conclusion, Torsvik et al. (1987) state that (quote): (1) «The principal magnetic signature in the HDM postdates or witnesses the final stage of a major tectonothermal event». (2) «The area has suffered postdepositional regional folding». (3) «The orogenic deformation of the Devonian rocks is related to a post-depositional N-S compressional event and is possibly part of a final phase of easterly nappe translation, the Svalbardian or Solundian Orogeny».

Large-scale tectonothermal processes and models, such as those mentioned in the above conclusions, are discussed in Ch. 6.

5.5.5 BEDDING-NORMAL CUMULATIVE STRATIGRAPHICAL THICKNESS

General

Kolderup (1925) was the first to give an estimate of the stratigrpahical thickness of the HDM, suggesting it to be *«more than 1000 metres»*. This estimate has later remained the only one available for the HDM in the published literature, and the figure has therefore been referred to in all subsequent papers. *In the following this estimate is shown to be wrong.*

The estimate of Kolderup (1925) was based on little more than the altitude of the highest mountain peak in the HDM. Apparently, a better estimate could not be given because the bedding orientation was not found due to the scarcity of bedding indicators in the massive conglomerates of the HDM. After Kolderup's (1925) pioner work, the Håsteinen massif has been the only Devonian massif in western Norway *not* subjected to renewed separate investigations, and the erroneous estimate of **1000 m** has therefore first been corrected during the cource of the present study.

The large degree of consistency and constancy in the dip of bedding both along the axial trace of the Osstrupen syncline (i.e. along the plunge of the fold axis), and also in the limbs of the syncline (Sect. 5.5.2.2), have made it possible to achieve good three-dimensional geometric control of the bedding orientation along the axial trace. These data show that the bedding along the axial trace has a fairly constant dip towards the east-southeast all the way (**Plate 2**). This geometrical organisation of the bedding allows trigonometrical calculations of the stratigraphical thickness of the deposits, i.e. the thickness measured orthogonal to the bedding.

Results

The W-E distance from the Devonian/substrate contact at Gravaneset, to the contact in the easternmost part of Svardal, is approximately **7.2 km**. The distance is measured horisontally, and at a level close to sea level. The strata are taken to dip **53°** towards ESE, which is the average plunge of the fold axis obtained from **376** bedding recordings aquired throughout the massif (**Fig. 5.51a, 5.51b**). This dip is considered to be representative, since the local fold axes at the plateau to the south of Gravanesholten, and at the Osen area, plunge **58°** and **50°**, respectively (Sect. 5.5.2.2). A dip of **53°** is close to the average between the last two values. Simple trigonometry, calculated from the expression

$$X = 7.2 \sin 53$$

as shown in **Fig. 5.73**, gives a stratigraphical thickness of

$$X = \mathbf{5.8\ km}.$$

This is a minimum estimate of the stratigraphical thickness of the massif, since two parts of the HDM have not been included; the part of the HDM situated east of the Vassetvatnet-Svardal valley, and the eroded part of the basin that existed west of the present HDM-outcrops. Based on the present data and a comparison with neighbouring massifs, however, it is possible to give a thickness estimate for the *entire* HDM, i.e. including the eastern part of the massif. When the Hornelen Devonian massif to the north of the HDM is considered, it is seen that the very consistent eastward dip of the strata along the axial trace is a very characteristic feature (Kolderup 1927a; Steel & Aasheim 1978; Seranne & Seguret 1987; and Norton 1987). It is thus very reasonable to assume that this is also the case for the *whole* of the HDM, particarly so since the dips of bedding in the *western* half is so consistent.

Based on the procedure used for the western half of the HDM, and assuming that the bedding continues with the same orientation in the eastern half, the horizontal W-E distance — which is about **14 km** for the entire HDM — gives the expression

$$X = 14 \sin 53$$

yielding a corresponding stratigraphical thickness of

$$X = \mathbf{11.2\ km}.$$

It is very likely that this estimate is close to the true bedding-normal cumulative stratigraphical thickness. *It is thus suggested that **11 km** from now be used as an estimate of the stratigraphical thickness of the entire Håsteinen Devonian Massif.*

Discussion

All of the eastward-dipping layers along the axial trace are lying with depositional contacts on top of each other, and therefore now define a stratigraphical column. All these layers were originally deposited (sub-)horizontally. It is important, however, to realise that all these layers have <u>not simultaneously</u> been lying horizontally on top of each other in a *vertical* stratigraphical coloumn – and have <u>not</u> later been tilted *en bloc* to achieve the present eastward constant dip. Instead, as will be further discussed later (Sect. 5.5.8 and Ch. 6), the stratigraphy was formed in an *extensional regime* that produced a basin located to the west of a westward-dipping listric fault that probably soled out in the subjacent top-to-the-west, low-angle and shallowly westward-dipping Nordfjord–Sogn Detachment Zone (see Sect. 2.8), implying that all new sedimentary layers, which were constantly deposited (sub-)horizontally in the eastern part of the basin during the on-going westward extension, were *successively rotated* into the eastward dip as they were transported down- and westwards on the fault. Such a rotational process allows large stratigraphical thicknesses (measured orthogonal to bedding) to be produced, even though the actual *vertical* sedimentary thickness at any time may have been considerably less. The thickness discussed here is therefore denoted "***bedding-normal cumulative stratigraphical thickness***".

The stratigraphical thickness of the HDM may be compared with neighbouring Devonian massifs. The Hornelen Devonian massif, with its **50–60 km** E-W length, has a cumulative stratigraphical thickness estimated to **20–25 km**. Also this thickness was originally calculated by Kolderup (1927a), who gave a correct estimate in this case. The bedding is oriented with a constant eastward dip of about **25 °**. Thus, the entire HDM, with its length of **14 km** and a thickness of **11 km**, has *half* the stratigraphical thickness of the Hornelen massif, with only ***1/4*** of the E-W length. The comparatively large thickness of the HDM is a result of the steep bedding-dip (**53°**) in the massif. The Kvamshesten massif has a cumulative stratigraphical thickness estimated to **7 km** (Seranne & Seguret 1987), and an average eastward bedding-dip of about **15–20°** (Seranne & Seguret 1987; Osmundsen et al. 1998). The Solund Devonian Massif has a cumulative stratigraphical thickness of **6 km** (Seranne & Seguret 1987).

The subject of performing restoration (back rotation) of the east-dipping bedding, to the (sub-)horizontal angle that existed at the time of deposition, reveals interesting geometrical features. As shown in Sect. 5.3, the Devonian/substrate contact of the HDM is clearly a ***primary depositional unconformity*** (despite local tectonic modifications), i.e. the HDM rests ***in situ*** with respect to the immediate ***substrate***. This situation is different from the situation in the Hornelen Devonian massif. In Hornelen, the east-dipping strata now rest against a shallowly westward-dipping ***brittle fault*** of possibly Mesozoic origin, which separates the Devonian rocks from the Devonian ***detachment zone mylonites*** below. Originally, however, the Hornelen sediments were, as mentioned above, deposited in a basin situated west of, and above a westward-dipping, basin-controlling listric fault. The sediments were thus deposited against this tectonic surface, and were successively rotated down the fault to receive their east-dipping orientation. "Back-rotation" of the Hornelen layers to their original subhorizontal orientation would simply imply a reversal of this process. Rigid back rotation of the Håsteinen layers, however, would have quite different implications: the subjacent unconformity stitches the HDM to the substrate, and "back rotation" would thus also rotate the ***substrate*** with the ***same magnitude***. As explained in detail later (Sect. 5.5.8 and Ch. 6), this shows that the "Hornelen type" depositional model, with deposition against a listric fault plane, cannot be applied to the Håsteinen area, where deposition occurred against a primary depositional unconformity. A new model is needed for the deposition in the HDM, and such a model is presented in Ch. 6.

As further discussed in Ch. 6, the magnitude of the bedding-normal cumulative stratigraphical thickness of the HDM is important, because it represents a minimum estimate of the amount of extension along the subjacent low-angle detachment zone. In addition, the ***amount of dip*** of the layering of the stratigraphy represents an estimate of the amount of rotation which occurred during the process that eventually formed the "constant-eastward-dip"-style of the bedding.

5.5.6 M_1-METAMORPHISM

General

The present section deals with the metamorphic mineral changes related to D_1 (M_1) in the HDM. Effects of earlier ***diagenetic*** processes will be shortly mentioned, but will not be subjected to separate treatment in the present work. The objective of the investigation has been to establish the overall metamorphic grade of the rocks, and the description of metamorphism is based on textural and mineralogical features in thin section.

The degree of mineralogical changes due to dynamothermal metamorphism is very limited throughout the HDM. This is the combined result of two major factors: (1) the relatively low temperature and pressure conditons which have prevailed during the deformation, and (2) the limited degree of D_1-related deformation which have restricted the grain-scale dynamic processes normally triggering metamorphic mineral changes.

In the HDM, the cleaved sandstones of the Gravaneset sandstone unit (Sect. 5.5.3.2) are the most suitable rocks for investigating metamorphic mineralogical phase changes, since metamorphic reactions will be most pronounsed in the the most deformed rocks (Yardley 1989). The degree of tectonically induced metamorphic recrystallisation in the sandstones is large enough to allow observation of individual metamorphic minerals. These are described in the following.

Chlorite

The amount of metamorphic chlorite in the quartzo-feldspatic sandstones of the Gravaneset sandstone unit varies, but may be up to **5–10 %**. Chlorite has occasionally grown as numerous minute grains in the otherwise sericitic microcrystalline matrix (**Fig. 5.74**). The mineral has also grown as individual larger grains, both within the general matrix, and more locally between adjacent clastic grains (**Fig. 5.75**). Most metamorphic grains are usually much smaller than the average clastic chlorites, and this provides a means of separating the two. In addition the clastic chlorites usually have sharply defined edges, whilst the metamorphic ones commonly are more irregular aggregates. Occasionally, however, it may be difficult to separate clastic grains from metamorphic ones. Chlorite has locally partly replaced other minerals, such as epidote (**Fig. 5.75**), and has also grown at the expense of feldspar (**Fig. 5.76**) and opaques. Chlorite has also grown as fibres from the ends of prismatic clastic chlorite grains.

The metamorphic chlorites tend to be oriented with their longest dimension parallel to the S_1-cleavage (**Fig. 5.75**). This indicates that they have grown syntectonically during the D_1-deformation. Another explanation for such a cleavage-parallel orientation could be that the chlorite has grown post-tectonically, mimicking the S_1-cleavage. This is, however, considered very unlikely, due to the lack of a cleavage that would be strong enough to control such post-tectonic growth, and to the fact that chlorite easily overgrows earlier fabrics when growing isostatically (Spry 1969). Late or post-tectonic chlorites overgrowing the S_1-fabric are not present.

Clorite may form as a ***diagenetic*** mineral (e.g. Winkler 1979), and some of the new-grown clorite might have originated during early diagenesis. However, the larger new-grown ones in the S_1-cleavage (**Fig. 5.75**) are interpreted to have been formed by metamorphic processes.

White mica

Up to **15 %** of the rock may locally consist of new-grown white mica / sericite, and this mineral is the dominating metamorphic mineral in the rock. White mica has grown in the form of sericite in the matrix between the clastic grains (**Fig. 5.68**). Occasionally, larger grains of white mica have been produced (**Fig. 5.68**), but these are still usually much smaller than most clastic grains of white mica (**Fig. 5.65**). White mica has grown at the expense of feldspar. The feldspar grains display all stages of alteration to white mica. In feldspar grains which are only slightly altered, the sericitisation may be confined to the grain margins (**Fig. 5.76**). The most altered feldspar grains may be completely or almost completely replaced by sericite (**Fig. 5.77**). In some feldspar grains the sericitisation process has developed gradually from the grain margins and inwards (**Fig. 5.76**), but in most grains the patches of sericite is distributed evenly throughout the grains (**Fig. 5.77**), indicating that the growth of sericite was initiated in all parts of the grains more or less simultaneously. White mica has grown in strain shadows, typically sheltered by quartz grains (**Fig 5.67**). Sericite has also formed as films along the contacts between quartz or feldspar grains. These films are parallel to, and thereby contributing to, the S_1-cleavage (**Fig. 5.65**). Subgrains of sericite have occasionally grown on the rims of clastic white micas which have thereby been partly replaced, and some clasic grains may be almost completely replaced by sericite.

From these observations it is concluded that the growth of sericite/white mica has occurred syn-tectonically during the D_1-deformation.

White mica may grow during ***diagenetic*** conditions (e.g. Yardley 1989). The growth of sericitic white mica in the matrix, and the sericitisation of feldspar, were therefore presumably initiated during diagenesis, but the large amount of new white mica grains which have grown parallel to the S_1-cleavage is interpreted to be a result of metamorphic processes.

Quartz

The clastic quartz grains have experienced subgrain development due to the Devonian dynamothermal metamorphism. The smallest clastic grains appear to have suffered the most extensive polygonisation, and might occasionally be almost completely replaced by tiny subgrains (**Fig. 5.64**) forming polycrystalline aggregates. Features of similar type have been described from quartzites by Spry (1969). The larger grains have developed subgrains that are commonly situated along the grain margins (**Fig. 5.64**, **5.67**, **5.68**). This suggests that the subgrains were initially formed at the margins, and developed further inwards towards the centre of the grains with increasing strain. Subgrains are often seen to be elongated parallel to S_1.

The majority of the larger clastic grains that were polygonised during the Devonian D_1-deformation have relict cores that are monocrystalline. This suggests that the grains were monocrystalline prior to the Devonian deformation. The unaffected interior of such minerals are frequently elongated in the S_1-cleavage direction. Almost all relict quartz grains display undulose extinction. This is a common effect from folding in quartzose sandstones (Spry 1969), such as those sandstones present in the Gravaneset sandstone unit. The undulose extinction is thus interpreted as an effect of the Devonian deformation, although grains with relict undulation inherited from the source area, might be present. Formation of subgrains has frequently occurred at the contacts between the quartz grains. Beard-like fibrous quartz has grown in strain shadows between larger quartz grains (**Fig. 5.65**). The fibres are oriented within, and are contributing to, the S_1-fabric.

The metamorphic recrystallisation of quartz described above is therefore syn-tectonic and related to the Devonian D_1-deformation.

Generally, quartz grains may experience initial stages of subgrain formation and pressure solution during deep ***diagenesis*** under increased burrial (e.g. Blatt et al. 1980), and such processes may have taken place in the HDM. However, following the criteria suggested by Turner (1981), the relative extensive polygonisation and parallel alignment of quartz grains into a prominent tectonic fabric indicate that diagenesis has been replaced by higher energetic metamorphism as the dominating mineral-transforming process.

Feldspar

As noted above, feldspars show alteration to sericitic white mica. The most extensively sericitised feldspar grains (**Fig. 5.77**) might originally have been K-feldspar or calcic plagioclase. Under low to medium grade metamorphic conditions, and with accessible H_2O, K-feldspar will generally undergo extensive sericitisation (e.g. Spry 1969). The albitic feldspar grains appear to have been only slightly altered, with growth of sericite essentially along the rims (**Fig. 5.76**). This indicates that albite was less unstable than the other feldspars.

Diagenetic processes are assumed to have played a role in the early stages of feldspar alteration.

Epidote

Clastic grains of epidote appear to show minor alteration to chlorite, but most grains are intact. Minor amounts of metamorphic epidote have possibly grown as a result of breakdown of feldspar.

Evaluation of metamorphism vs. diagenesis

The HDM has been subjected to limited tectonometamorphic alteration, and the distinction between metamorphic and diagenetic effects is thus important. According to the criteria suggested by Turner (1981), the overall texture in the Gravaneset sandstones, with the combination of mineral transformations and a relatively prominent ***cleavage***, clearly indicate that pressure and temperature had exeeded well beyond the diagenetic conditions, and that metamorphic processes were operating. Although the minerals described above are not themselves precise metamorphic index minerals, and therefore not diagnostic for a particular and well constrained P/T interval, information on the metamorphism may be obtained.

Metamorphic grade of the M_1-metamorphism in the HDM

The prograde metamorphism in the HDM is characterised by growth of e.g. chlorite, white mica, quartz, and possibly epidote, associated with cleavage development, as described above. The metamorphic mineral-changes described have been shown to be closely related to the D_1-deformation, and are accordingly termed M_1. The mineral textures and mineral transformations indicate that the sediments in the HDM have experienced one phase of prograde metamorphism which altered the sediments from an unmetamorphic state to the present metamorphic condition. The prograde metamorphism has evolved as a result of gradually increasing overload which has subjected the sediments to increasing temperature and pressure. Metamorphic mineral reactions have been facilitated by the D_1-related dynamic effects.

The metamorphic minerals growing syn-tectonically in the S_1-cleavage during the M_1-metamorphism indicate that the HDM has experienced P/T conditions corresponding to the upper anchizone/epizone (prehnite-pumpellyite facies of Yardley 1989) or possibly lower greenschist facies. Such a degree of metamorphism would be compatible with data from the other Devonian massifs in western Norway (see below).

Low-grade index minerals with better constrained P/T-ranges, such as e.g. *prehnite, pumpellyite, actinolite, stilpnomelane or pyrophyllite,* have not been observed. Since growth of the three first of these minerals preferentially occurs in more "basaltic" bulk chemistries (Yardley 1989), and stilpnomelan in more pelitic/basaltic environments, (Yardley 1989), their absence in the quartzo-feldspatic sandstones of the Gravaneset sandstone unit is probably due to unfavorable bulk chemical conditions. *Pyrophyllite* might be present, but its close microscopic resemblance to sericite (Mackenzie & Guilford 1980), and small grain size, make separation from white mica impossible under the microscopic.

The M_1-metamorphism, which is prograde with respect to pressure and temperature, caused retrograde mineral-reactions in some clastic grains and minerals. These retrograde reactions are characterised by the breakdown of especially clastic feldspar, but also possibly of epidote. In addition, clastic quartz grains were replaced by quartz subgrains. This shows that the clastic grains, from their source area, were adjusted to more elevated P/T conditions.

P/T considerations for the HDM

Although exact temperature and pressure estimates cannot be given, the data allow some rough estimates. From the metamorphic and related *tectonic textural development* of the HDM it is reasonable, following Turner (1981) and Yardley (1989), to suggest that the temperature reached a level of **300 +/- 50 °C**. This estimate is also supported by P/T-investigations from other west Norwegian Devonian massifs (see below).

If a moderate heat flow with a thermal gradient of **25 °C/km** is assumed to have been present in the HDM, this indicates an overburden with a thickness of **12 +/- 2 km**. If a normal crustal rock density of **2.8 g/cm^3** is assumed for the sediments, the pressure gradient will be about **1 kbar** per **3.6 km** rock column. This indicates that pressures may have reached **3.3 +/- 0.5 kbars** in the Devonian rocks. If we assume a high heat flow and a thermal gradient of e.g. **40 °C/km**, an overload with thickness **7.5 +/- 1.3 km** is obtained. With the same pressure gradient as above, the pressure would have been about **2.1 +/- 0.4 kbars**. In Basin & Range, thermal gradients as high as **50 °C/km** has been reported (REF ?). Such a thermal gradient would imply an overburden of **6 +/-1 km** and a pressure of **1.7 +/- 0.3 Kbar**. However, the thin and hot Basin & Range crust is anomalous and not representative for the western Norway Devonian crust.

Comparison with other Norwegian Devonian massifs

P/T estimates from metamorphic index minerals (Solund): Seranne & Seguret (1987) have studied the metamorphism in the *Solund Devonian Massif*, and reported metamorphic growth of the minerals prehnite, zoisite, clinozoisite, chlorite, albite, quartz and possibly pumpellyite. They concluded that the sediments have experienced a temperature of at least **200–300 °C**, corresponding to the prehnite-pumpellyite facies. However, they also reported the occurrence of metamorphic actinolite, indicating temperatures in the range of **300–350 °C**, corresponding to the beginning of greenschist facies. Furthermore, they concluded that the mineral assemblage is of a typical low-pressure type, suggesting an overburden pressure of less than **3 kbars** (depth less than **11 km**). The data and results of Seranne & Seguret (1987) support the temperature estimate given above for the HDM.

Temp. estimates from vitrinite reflectance (Hornelen, Kvamshesten, Ytre Byrknesøyer): Temperature estimates relevant to the HDM have also been obtained from the Hornelen Devonian Massif. The Hornelen Massif is situated less than **10 km** to the north of the HDM. The temperature data from Hornelen are based on analyses of reflectance of humic organic matter on shales from the easternmost part of the massif (carried out by Michelsen 1992). A temperature of about **250–300 °C** is suggested from these analyses (Michelsen 1992).

The Hornelen Devonian Massif has probably experienced P/T conditions not very different from the conditions in the HDM, and the temperature for the Hornelen Devonian Massif of **250–300 °C** therefore supports the estimate of **300 +/- 50 °C** suggested for the HDM.

The vitrinite reflectance technique was used to analyse whether the heating of the Devonian sediments at one specific localitiy occurred rapidly or slowly compared to another locality (Michelsen 1992). Hense, a comparison was made between the data obtained from the eastern part of the Hornelen Massif and corresponding data from the western parts of the Devonian deposits in the areas of Kvamshesten and Ytre Byrknesøyer. This comparison showed that the heating rate in the *eastern* parts of the West Norwegian Devonian massifs has been considerably higher than the rate in the *western* parts. This is compatible with the detachment model which states that high temperature eclogitic rocks were uplifted just to the east of the Devonian basins (Sect. 2.11), thus providing faster heating in these areas.

P/T estimates from Si^{4+} content in white mica, and illite crystallinity (Smøla): Atakan (1988) and Bøe et al. (1989) used *microprobe* analyses of phengitic white mica (from sandstones) in combination with *XRD* analyses of the < 2μ fraction (from silts and very fine-grained sandstones) to establish an estimate of the P/T conditions during the main metamorphism in the *Smøla Devonian deposits*, which are situated in the coastal areas of the Møre-Trøndelag region. The *microprobe* analyses yielded information on the *Si^{4+} content* in the new-grown mica crystal lattice. Velde (1967) showed that the Si^{4+} content in phengitic white micas is dependent on P/T relations, and that the Si^{4+} content therefore may be used for P/T estmations. Based on this, Velde (1967) empirically established stability curves in a P/T diagram for different amounts of Si^{4+} (**Fig. 5.78**). The metamorphic white mica at Smøla had a Si^{4+} content corresponding to **3.42** (**Fig. 5.78**). The *XRD* analyses of Atakan (1988) and Bøe et al. (1989) yielded information on the *illite crystallinity index*. This method utilises the fact that the crystallinity of the illite generally increases as a function of the temperature. The result of the illite crystallinity analyses was that the crystallinity corresponded to a temperature interval of **300–350 °C**. Atakan (1988) and Bøe et al. (1989) combined this temperature interval with Velde's (1967) P/T diagram of stability curves for different amounts of Si^{4+} in a mica lattice, and obtained a corresponding pressure of **4.1–4.8 kbars**.

Discussion: With a rock density of **2.8 g/cm^3**, this pressure corresponds to a crustal depth of **14.8 –17.3 km**. The pressure is surprisingly high, and it may be questioned whether the Si^{4+} contents in the mica lattices have really been in equlibrium with the surrounding material. A fundamental prerequsit for the method will of course be that the exchange of the Si^{4+} between the phengites and the surrounding minerals is continuous, so that the Si^{4+} content really reflects the P/T conditions. However, Atakan (1988) and Bøe et al. (1989) paid no attention to the possibilities of equilibrium problems, and the validity of the pressure estimates may thus be questioned. The temperature, however, was obtained independently, and the estimates of **300–350 °C** agrees well with the above temperature estimates for the HDM.

P/T estimated from vein inclusions (Hornelen, Kvamshesten, Solund): Svensen & Jamtveit (1999) and Svensen et al. (2001) studied veins and associated inclusions, as well as authigenic minerals of the wall rock, in the Hornelen, Kvamshesten and Solund massifs. Vein compositions were used to estimate temperatures during crystallisation of the veins. In the Solund massif, the presence of biotite was taken to indicate temperatures exceeding **300 °C**, and chlorite thermometry gave temperatures in the range **305–330 °C**. In other words, the authors thus estimated the temperature to have been **315 +/- 15 °C**. The Hornelen and Kvamshesten massifs were interpreted as having experienced a temperature of **250 +/- 20 °C**, based on presence of epidote, and absence of amphibole. The temperature estimate of **315 +/- 15°C** for Solund and **250 +/- 20 °C** for Hornelen and Kvamshesten were then combined with fluid inclusion isochores from these veins, giving pressures of **3.4 +/- 0.2 kbar** (1 bar = 105 Pa) and **2.4 +/- 0.4 kbar** respectively. Using a linear pressure gradient, and a mean rock density of **2.6 g/cm^3**, these pressures were found to correspond to a depth of burrial of **13.4 +/- 0.6 km** for

Solund and **9.1 +/- 1.6 km** for Hornelen and Kvamshesten. The formation of veins were interpreted as a result of burrial and local hydrothermal fluids, i.e intrabasinal processes.

Discussion: The pressure estimates of Svensen and co-workers are substansially lower than those of Atakan (1988) and Bøe et al. (1989) referred above, and lies within a range that is geologically reasonable for a scenario with burrial in a basin. The Solund pressure estimate is high, although not geologically impossible, and the Hornelen/Kvamshesten estimate is moderate. The temperatures are in accordance with those suggested above for the Håsteinen Devonian Massif.

Comparison between Solund and Smøla: Sturt & Braathen (2001) investigated the structural and tectonometamorphic development of mainly western parts of the Solund Devonian Massif, and reported that the metasediments were strongly folded and cleaved. Thermobarometric methods were not a part of the study, but in thin-section they observed that the cleavage were "***decorated by phengite and chlorite***". Other minerals defining the cleavage were chlorite, quartz, epidote, carbonate, opakes and subordinate biotite. From this, the authors concluded that the deposits had been subjected to greenschist facies metamorphism. The deformation and metamorphism were interpreted as "reminiscent of the tectonometamorphic development of the Devonian rocks of Hitra and Smøla (Bøe et al. 1989)", i.e. suggesting the same high-P /moderate-T conditions as referred above (**4.1– 4.8 kbars** = **14.8 –17.3 km**; **300–350 °C**).

Sturt & Braathen (2001) also discussed the work carried out by Svensen & Jamtveit (1999) on the veins and authigenic minerals of the Solund Massif, and disagreed with the interpretation that these features were a result of intrabasinal processes. Sturt & Braathen (op. cit.) instead argued that the veins and new-grown minerals were a result of regional deformation and metamorphism related to their Solundian Orogeny.

Discussion: The suggestion by Sturt & Braathen (2001) that the Solund Massif experienced the same high pressures (**4.1– 4.8 kbars** = **14.8–17.3 km**) as those claimed for the Hitra and Smøla deposits, raises the same questions as above, regarding the validiy of the pressure estimates. This becomes particularly pertinent in the light of the work by Seranne & Seguret (1987), where the metamorphic mineral assemblage of Solund is interpreted as a typical low-pressure type, suggesting an overburden pressure of less than **3 kbars** (depth less than **11 km**). Concerning the temperature, the use of the Hitra–Smøla estimate of **300–350 °C** for Solund agrees well with the estimate by Svensen and co-workers of **315 +/- 15 °C** for Solund, which is again compatible with the estimate for the Håsteinen Devonian Massif in the present thesis.

From the above review it appears that the temperature of **300 +/- 50 °C**, which is suggested, in the present thesis, as the maximum temperature in the HDM, is a resonable estimate.

5.5.7 FAULTS AFFECTING THE HDM

5.5.7.1 GENERAL

Section 5.5.7 gives an overview of the faults affecting the HDM. The purpose is not to give a full analysis of the faults, but to present some preliminary observations and interpretations.

Unless otherwise stated, the term "fault" is generally taken to cover both semiductile shear zones and brittle faults. A number of individual faults will be described, and for each of them, the section will (1) briefly mention the fault rocks, if present, (2) elucidate the orientation of the faults and (3) analyse the *amount of displacement* on the faults. The latter is the main purpose of the section, in order to establish whether or not the faults are significant for the 3-D architecture of the HDM, which is discussed in the next section (Sect. 5.5.8). In addition to this, some general aspects of the relationship between fault rock development and the amounts of displacement shall be presented. Where the amount of displacement *cannot* be directly observed due to lack of "off-set relationships", an indication of the amount of displacement may be inferred by comparing fault rocks that are present in faults with *known* displacement, with fault rocks present in faults with *unknown* displacement.

The faults in the HDM are divided into two categories; those cutting *through* the HDM (Sect. 5.5.7.2), and those present along the *margins* of the massif (Sect.5.5.7.3). Emphasis will be put on the faults crosscutting the HDM, since these allow observation of the amounts of displacement (see below). The faults along the margins will be given only a brief treatment. A summary is given at the end of the section (Sect. 5.5.7.4).

5.5.7.2 FAULTS CUTTING THROUGH THE HDM

General

The HDM is affected by a large number of faults and joints. The faults may be of a *prominent* or *less prominent* type, both of which may *reach* the marginal Devonian/substrate contact, or the faults may be of a type *not reaching* this contact. The faults presented are *numbered* (1-11) clockwise around the massif according to their geographic position and regardless of type of fault, starting in the west-northwest. An overview of fault-locations and fault-numbers is shown on the sketch-map of **Fig. 5.79.** Detailed information on the various faults may be obtained from **Plates 1/2/4/5** (and **Plate 3/App. A** if desirable), as appropriate in each case. (Note that fault the numbers are *not* indicated on **Plates 1/2/3/4/5** and **App. A**).

Some of the *prominent* faults, defining prominent lineaments on aerial photos and topographic maps, as well as in the field, may be traced for **several km** across the massif, and some of these faults will be briefly described in the following (**Fig. 5.79**). The faults displaying the most prominent topographical signatures have been placed on the main map (**Plate 1**). These are (with fault numbers in brackets): (2) the Osen–Mannen(-west) fault, (3) the Osen–Mannen(-east) fault, (4) the Vassetbrekka–"Galmannsskåra" fault, (5) the Vassetbrekka–Stegabruna fault, (6) the Leirvågfjellet fault, (7) the Osen–Teigafjell fault, and (8) the Breiskåra–Østre Håsteinen fault.

In addition to these faults, the topographically *less prominent* faults at the westernmost Gravaneset and Straumsnes areas will be included in the following treatment (**Fig. 5.79**). These faults are shown on **Plate 1**, **Plate 4** and **Plate 5**. Although these faults have limited lateral extension when judged from the topographical signature, the faults cut the marginal Devonian/substrate contact. These faults are (with fault numbers in brackets); (1) the Straumsnes fault (**Plate 1**), and (11) the group of three faults at Gravaneset: the southern (11a), central (11b) and northern (11c) Gravaneset fault (**Plate 4**, **Plate 5**), of which fault 11a and 11c will be treated.

All the faults listed above, whether prominent or less prominent, can be traced from an intra-Devonian position and out to where they reach (and cut) the marginal Devonian/substrate *contact*. A few faults which *cannot* be followed to the point where they cut the Devonian/substrate contact, will also be included (**Fig. 5.79**). These faults are (with fault numbers in brackets); (9) the group of faults on the plateau to the south of Gravanesholten: the southern (9a), central (9b) and northern (9c) Gravanesholten-plateau fault (**Plate 4**), of which fault 9a near the Osstrupen syncline axial trace will be treated, and (10) the semiductile fault in the road-cut in the far west about **10 m** to the north of the inferred position of the axial trace of the Osstrupen syncline (see **Plate 4**).

Method

As mentioned above, the main purpose of the discussion of faults is to elucidate the amount of displacement on the faults. To be able to test the amount (and possibly direction) of displacement, it is crucial that the faults cut some sort of structural *marker* which reveals the off-set, since the faults themselves are rarely exposed. In the HDM, the presence of massive debris-flow conglomerates, and, hence, lack of laterally persistent horizons such as sandstone layers, implies that no such markers are present within the HDM itself. The only place where the amount of displacement may be estimated is at the *margin* of the HDM where the faults cross the contact between the Devonian rocks and the substrate, or at the contacts between the HDM and the inliers, notably the Høgdene inlier. (Most faults meet the Devonian/substrate contact in at least one lateral direction). In the following, the faults are therefore divided into *two groups*; (1) faults *suitable* for test of displacement (i.e. cutting the marginal Devonian/substrate contact at a reasonably high angle), and (2) faults *not suitable* for test of displacement (i.e. not reaching the contact, or bending into parallelism with the contact).

Faults suitable for test of displacement

For the following faults, the amount of displacement can be tested in the field:

- *Fault 1: The Straumsnes fault* (**Fig. 5.79**) is well exposed in the area where the fault cuts the marginal Devonian/substrate contact. The offset of the contact yields a fault- displacement of maximum **50 m** as measured in map view (**Plate 1**). This brittle fault has a cohesive flint-hard (hereafter termed “flinty”) cataclastic fault rock which is about **10 cm** thick. The colour of the fault rock is partly grey and partly yellow-green, indicating that material from quartz-epidote veins were involved in the fault rock. The cataclastic nature of the fault rock indicates that the movement probably occurred post-HDM-D_1. The fault is oriented **142/52 SW** (**Appendix A: map no. 5e**), i.e. with a moderate to steep dip towards SW.

- *Fault 2: The Osen–Mannen(-south side) fault* (**Fig. 5.79**) is not exposed at the marginal Devonian/substrate contact. However, based on adjacent wall rock exposures, the fault appears to have a displacement of maximum **20 m** measured in map view (**Plate 1**). The fault has not been followed in the field, but as indicated in the western profile, a steep dip towards SSW is inferred (**Plate 2**). This dip is based on the interaction of the fault trace and the topography: On the main map (**Plate 1**) it is seen that the fault trace defines a *convex* "arch" across topographical *depressions* (viewing NE), a *concave* "arch" across topographical *highs*, and a straight line across largely planar topographical surfaces; i.e. yielding the inferred dip. This fault trace behaviour also implies that the fault surface is roughly *planar*, i.e. with a fairly constant strike and dip. As we shall see, not all fault surfaces in the study area are planar.

- *Fault 3: The Osen–Mannen(-north side) fault* (**Fig. 5.79**) is not exposed at the intersection between the fault and the Devonian/substrate contact in the northwest, but maximum displacement in map view is of the order of **10 m**. However, in the contact area it cannot be completely excluded that the fault curves into and follows the Devonian/substrate contact towards the west instead of going straight into the substrate. In such a case the displacement may be larger. The fault has not been followed in the field, but aerial photos show the complex anastomosing fault pattern, of which the main features are shown on the map (**Plate 1**). Assuming a planar fault surface, (which is a reasonable approximation since the fault coalesce with the planar *Fault 2* in the southeast direction towards Osen), means that a vertical to steep SW dip is inferred from the interaction of the fault trace and the topography, as illustrated in the western profile (**Plate 2**).

- *Fault 4: The Vassetbrekka–"Galmannsskåra" fault* (**Fig. 5.79**) has, in the Devonian rocks, been traced in the field all the way from the Devonian/substrate contact at Vassetbrekka (east) to the "Galmannsskåra" area (west). The fault is crossed by both the western and the central profiles (**Plate 2**). In the west, at "Galmannsskåra", the fault splits into three branches across the Høgdene inlier (**Plate 1**). The *northernmost* of these faults follows the cleft of "Galmannsskåra" which goes from the Vikafjell plateau in the southeast to the base of the mountainside in the northwest (**Plate 1**). This branch forms the major fault trace, but the fault rock itself is not exposed. Displacement along this major branch appears to be less than **10–20 m** when judged from surrounding exposures at the southern margin of the inlier. In the eastern parts, at Vassetbrekka, the fault appears to follow parallel to the Devonian/substrate contact, and possibly has contributed to the position of the contact. Although the amount of displacement is here slightly more uncertain due to lack of structural markers, it is reasonable to assume a small displacement also in this area, as illustrated in the central profile (**Plate 2**). To the east of the Vassetbrekka area, the continuation of the fault is uncertain.

An exposure of fault rocks is found on the *southernmost* of the three branches crossing the Høgdene inlier, at a locality situated in the Devonian rocks 20 m south of the southern ("upper") contact of the inlier. Since exposures of these most prominent faults are very rare within the Devonian massif, the field-appearance of the fault rock will be briefly described, although displacement is negligible. The fault movement has effected the rocks in a total width of about **60 cm** (**Fig. 5.80a**). Most of the fault zone is defined by fracturing, although the rocks are fully cohesive. In the centre of the zone, a flinty greenish-yellow ultra-cataclastic cohesive fault rock with a thickness of only **1–3 mm** is present (**Fig. 5.80b**). The fault may thus be classified as semibrittle to brittle. Fault 4, which is one of the most prominent faults in the HDM in terms of topographic signature, has thus produced only a very modest displacement fabric in a major fault splay. The amount of displacement along this southernmost branch of the Vassetbrekka–"Galmannsskåra" fault may be tested where the fault cross-cuts the Høgdene inlier, and the displacement is at the most **2 m** in the horizontal plane. It is theoretically possible that the finite displacement vector is parallel to the intersection-line between the fault plane and the Devonian/inlier-contact so that displacement may be present without visible off-set, but this is not considered very likely.

The orientation of the fault will be given closer attention. Towards the *east*, the strike/dip recordings of the fault surface is **111/67 SSW** at the eastern part of the Vikafjell plateau (**Plate 1**), and **114/56 SSW** at Vassetbrekka (**Appendix A**), as illustrated in the central profile (**Plate 2**). In the *western* parts of the fault, around the upper part of the cleft of "Galmannsskåra", the *orientation* of the southernmost one of the three branching faults has been measured at the locality **20 m** south of the southern Devonian-inlier contact, coming out at **121/59 SSW**. A similar orientation is most likely also present in the two other branches, as illustrated in the western profile (**Plate 2**). Although the strike/dip orientation in the western part of the fault has changed slightly compared to the eastern part, the fault surface must be said to be essentially planar over this distance. On the northern mountainside, to the *west-northwest* of the **121/59 SSW**-locality, it is seen, on the main map (**Plate 1**), that the fault-trace curves to the *right* when looking west-northwestward from Vikafjell along the fault (i.e. curves from a W-E / WNW-ESE direction on the plateau of Vikafjell, to a WNW-ESE / NW-SE direction on the mountainside). In a fault with planar fault surface, however, a continuation of the fault orientation **121/59 SSW** west-northwestwards into the mountainside, should, because of the topography, have made the fault-trace curve into a *westward* direction, i.e. to the *left* when looking west-northwestward. However, the fault does not have this curvature in the mountainside; in fact the fault-trace curvature is “opposite” of what should be expected. The presence of this "opposite" curvature on the fault-trace shows that the fault is *not* planar in this area.

Viewed separately, the "opposite" fault-trace curvature in the mountainside could have been mistaken for indicating a NNE dip in this area. However, with a stable SSW dip documented for the *rest* of the fault, such a NNE dip for this area is *not* possible. This is because a NNE dip would give a fault surface with a *twisted* shape between east and west, and movements along such a fault would be impossible since large open gaps or a very wide fault zone would be created. The most likely explanation for the "opposite" curvature is that the fault *generally* continues with its SSW dip, but that the fault surface in the "Galmannsskåra" area contains a local "bend". Such a "bend" may have the shape of a cylindrical "fold", where the "hinge line" will be parallel to the displacement vector of the fault. The "bend" changes the strike orientation of the fault *surface* from a WNW-ESE to a NW-SE direction, producing the fault-trace with the "opposite" curvature.

- *Fault 5: The Vassetbrekka–Stegabruna fault* (**Fig. 5.79**) has, in the Devonian rocks, been followed in the field from the Devonian/substrate contact at Vassetbrekka (east) to the steep mountainside at the Høgdene inlier (west). Based on the very limited offset of the Devonian/inlier contact in the west, the amount of movement is apparently **some tens of metres**.

The orientation of the fault will again be discussed in some more detail. In the *east-southeast*, a fault-orientation measurement of **122/64 SSW** – and the interaction of topography and fault-trace – show that the fault dips steeply SSW, as shown in the central profile (**Plate 2**). In the mountainside to the *west-northwest*, the dip of the fault is again more uncertain. As seen on the main map (**Plate 1**), the fault-trace continuous as a fairly *straight* line from the plateau of Vikafjell and west-northwestwards down the mountainside to Stegabruna. Similar to Fault 4, a continuation towards the westnorthwest of the fault orientation found in the *east-southeast*, should, in a fault with a planar fault surface – and due to the effect of the topography/fault-trace interaction – have made the fault-trace curve markedly towards *west* in the montain side, instead of continuing as a straight line towards the west-northwest. The straight line shows that that the fault-surface of Fault 5 is not planar in this area. Viewed *separately*, the straight fault-trace down the mountainside could have been interpreted to show that the fault was generally vertical in the area. However, with a dip of **64°** ***SSW*** in the east-southeastern area, the fault cannot change to a vertical orientation in the west, since this would again produce a *twisted* fault surface, which cannot move very easily. Analogous to the situation in Fault 4, it is therefore most likely that the dip of **64 ° SSW** is the *general* dip of the fault, but that the fault-trace towards Stegabruna becomes straight (and the fault-dip steeper) as a result of a local *"bend"* in the fault surface in the mountainside to the westnorthwest. The fault has accordingly been drawn slightly steeper at Stegabruna (western profil, **Plate 2**) than at Vassetbrekka (central profile, **Plate 2**).

At Vassetbrekka, the fault has not been followed eastwards from the Devonian- substrate contact, but based on interpretation of aerial photos and the measured dip of the fault plane at 64° SSW, the fault appears to leave the Devonian/substrate contact. The eastward continuation is uncertain. No exposures of fault rocks were observed.

- *Fault 7: The Osen–Teigafjell fault* (**Fig. 5.79**) has been followed, in the field, all the way from Osen (west) to the mountain plateau of Teigafjell (east), and further eastwards along the *northern* side of the Teigafjell Substrate Wedge (TSW) from the plateau area and downwards on the steep eastern mountainside, as far down as it is accessable until the vertcal cliffs are reached. Likewise, the fault has been approached from the eastern side, and investigated below the vertical cliffs. From the **300 m** contour and downwards to lake Vassetvatnet, the fault (coming from the northern side of the TSW) is unexposed. Also the *southern* contact of the TSW is tectonised, and this contact has been traced in the same manner. Along this contact, exposures are found locally from Teigafjell to the shore of Lake Vassetvatnet.

On the main map (**Plate 1**) and in the profiles (**Plate 2**), the Osen–Teigafjell fault is drawn along the *northern* side of the TSW, as a *dip-slip fault* with minor reverse displacement. However, the possibility also exists that the fault could run along the *southern* side of the TSW, possibly as a *strike-slip fault*, again with minor displacement — and both these fault trace possibilities will be discussed in the following. Anyway, and regardless of the interpretation of fault trace alternative and displacement direction, the main message to be conveyed is that the movements are only minor.

In the Osen area, exposures of the fault are absent due to the extensive Quaternary deposits (**Plate 1**). The Osen–Teigafjell fault might possibly have a northwestward continuation in the faults going from Osen towards Mannen (Faults 2 and 3), but this cannot be confirmed due to the Quaternary cover at Osen. Eastward from Osen, the fault is exposed several places in the river along the northern margin of the Osen inlier. Along this margin, the fault is defined by fracture planes which may constitute a zone up to about **1 m** wide, but which is not so intensely developed that a distinct fault rock has been produced through the zone. The affected rocks are cohesive. At one locality near the central part of the northern contact of the inlier (UTM 1075 2955), the fault has produced a cohesive ultra-cataclastic flinty yellow-green fault rock that is **1–2 mm** thick. On the western part of the Teigafjell plateau, where the fault crosses the Devonian deposits that cover the substrate rocks in the area between the Osen inlier and the Teigafjell Substrate Wedge (TSW), no exposures of the fault are present, and only a depression markes the trace of the fault.

Eastwards on the Teigafjell plateau, the fault meets the Teigafjell Substrate Wedge (TSW), which has got tectonic contacts both along its northern and southern side. The fault coming fram Osen may thus continue either on the north or south side of the TSW. The interpretation as to which of these two faults that represent the eastward continuation of the Osen–Teigefjell fault imposes controle on whether the fault can move in the dip-slip or strike-slip direction. Both interpretations will be discussed in the following:

<u>Fault on the northern side of the TSW</u>: On the gently sloping part of the Teigafjell plateau, the fault surface is not exposed along the northern margin of the substrate wedge. East of this plateau, on the steep slopes of the mountainside continuing down towards the Al-lia area (above lake Vassetvatnet, **Plate 1**), a cohesive flinty yellow-green fault rock, up to **3 cm** thick, may be observed along the northern fault margin of the TSW.

Several measurements of the orientation of fault surfaces and fabrics have been recorded between Osen and the slopes located to the east of the top-flat of the Teigafjell Substrate Wedge (**Appendix A: map no. 10b, 10c, 11a**), and all measurements show steep dips to the S. This is illustrated in the central and eastern profiles (**Plate 2**).

Based on the generally very thin fault rock development along this fault, compared to elsewhere in the HDM where the displacement can be positively determined, it is reasonable to assume *a <u>very modest displacement for this fault, i.e. probably not more than about 50 m</u>*. The dip of the fault has been determined at

61° S at the central profile (**Plate 2, Appendix A: Map no. 10b and 10c**) and about **50° S** in the eastern part (**Plate 2**, **Appendix A: map no. 11a**).

Since the strike direction of the fault changes markedly on the top of Teigafjell (**Plate 1**), it is inappropriate to assign a pure strike-slip component to the fault. This means that the movement largely must have had a dip-slip character. Since substrate rocks are situated "on top" of Devonian rocks in an apparent "hanging wall" position ("older on younger") along the northern margin of the TSW, the fault appears as a small *reverse* fault, probably with a displacement in the order of less than **50 m**, as inferred above. This is illustrated in the central and eastern profiles (**Plate 2**).

The fault segment between Al-lia (just west of Lake Vassetvatn) and Teigafjell can be viewed on the main map (**Plate 1**). At Al-lia, the Devonian/substrate contact, which on the main map (Plate 1) is seen to come from the north, approaches the fault at an altitude of roughly **300 m**, and at Teigafjell the contact continues southeastwards at an altitude of roughly **500 m**. The main map (**Plate 1**) thus appear to give the impression that the fault has displaced the south block upwards with an amount of **200 m** on the vertical component, instead of the above estimated maximum of **50 m**. The apparent **200 m** vertical displacement on the map is, however, not real, but merely a *visual effect* resulting from the following situation: the overall Devonian/substrate contact surface in the whole area (mountainside) along the western side of lake Vassetvatnet has been found to dip steeply towards Vassetvatnet, i.e. forming the mountainside of a *<u>sub-Devonian</u>* "substrate mountain" called the "sub-Devonian Teigafjell" (Sect. 5.4.4.2). With the Devonian rocks forming a thin "carpet" on the steeply eastward dipping Devonian/substrate contact (Sect. 5.4.4.2), the **50 m** *real* upward movement of the TSW has allowed subsequent erosion to simply remove the thin Devonian "carpet" on the whole slope-surface of the Teigafjellet Substrate Wedge, thus giving the **200 m** *apparent* vertical displacement on the map.

<u>Fault on the southern side of the TSW</u>: In this alternative, the fault is interpreted to continue along the southern side of the Teigafjellet SubstrateWedge (TSW) and further southeastwards down to the shore of lake Vassetvatnet (**Appendix A: map no. 11a, 11d, 11e**). The fault trace on the Teigafjell plateau will then continue in its SE direction, without the marked change in fault trace direction described above of the “north side” alternative. The Devonian/substrate contact going from the southern side of the TSW to lake Vassetvatnet has experienced displacements, and has been defined on the main map (**Plate 1**), as a “tectonically modified primary unconformable contact”. However, in the present discussion, the possibility of the contact beeing a fault is explored.

Placing the fault on the south side of the TSW offers a different solution to the geometrical problem of explaining the presence of the Teigafjellet Substrate Wedge (TSW) as well as the Osen inlier. When the tectonic contact south of the TSW is interpreted as a fault, *<u>a dextral strike-slip movement along this fault could explain the outcrop pattern</u>*. The affects of this may be illustated from the Osen inlier area and the Teigafjellet Substrate Wedge area:

North of the Osen inlier, it is assumed that the Devonian sediments rest on the substrate with a depositional contact. If we focus on the average palaeo-topographic surface below the Devonian sediments, the overall depositional contact probably dips westwards in the same manner as the present-day average surface of the Osen inlier. A dextral strike-slip movement along the Osen–Teigafjell fault would then effectively lower the sub-Devonain overall depositional contact surface on the northern side of the fault, thus bringing the conglomerates in fault contact with the Osen inlier. Regarding the Teigafjellet Substrate Wedge (TSW), the dextral movement of this northern block would move the entire substrate wedge towards the east compared to the Devonian rocks to the south of the fault. This would facilitate the recent erosional removeal of the thin carpet of Devonian sediments on the wedge, leading to its present exposure. Within the shear zone between the TSW and the lake Vassetvatnet, shear sense indicators suggesting such dextral movements have in fact been observed, at a position **~280 m** above sea level.

The interpretation that the Osen–Teigafjell fault follows on the south side of the TSW, would require a new explanation for the fault on the *northern* side of the wedge. Along the southward-dipping northern fault, the TSW is lying on top of Devonian conglomerates. If this juxtaposition was not accomplished by reverse faulting, it could have been the result of minor sinistral strike-slip faulting. Such displacements would move the TSW and the above-lying Devonian sediments towards the east relative to the northern Devonian block, and thus bring the Devonian rocks to lie below the TSW rocks. Since the present-day carpet of Devonian sediments resting on the eastward-sloping side of the “Sub-Devonian Teigafjell” is generally thin, the resent erosion has brought the wedge into day-light. Striations or other shear direction indicators have not been observed along the fault on the north side of the TSW. If we trace the fault on the northern side of the TSW westwards, it meets the southern side Osen–Teigafjell fault. If the northern fault had *transected* the southern Osen–Tegafjell Fault, the development of the northern fault could be viewed as a separate case. This would allow a different interpretation of the north side fault, as the Håsteinen massif contain other faults having fault traces in the E-W to SE-SW direction that appear to belong to a different set of faults. However, in the westward direction, the northern fault rather appears to merge with the southern fault (the Osen–Teigafjell Fault). This suggests that the faults may be connected in some way.

The northern contact of the Osen inlier is, for considerable distances, *not* defined by the fault, as the fault is also to a large degree located *within* the Devonian sediments (**Plate 1**, and **Appendix A: maps no. 10b and 10c**). The fact that the northern contact between the Devonian and the inlier is not solely a fault, is compatible to the assumption that only minor movements have occurred along the fault.

In summary, the above discussion serves to illustrate some geometrical challenges conserning the interpretations of the faults bordering the Osen Inlier and the Teigafjellet Substrate Wedge (TSW). As mentioned above, the Osen–Teigafjell fault has, on the main map (**Plate 1**), been drawn on the *north* side of the TSW, and in the profiles (**Plate 2**), the fault is shown as a minor reverse fault. However, regardless of this, and as illustrated above, the important point to notice is that the movements have been only minor.

- *Fault 11a: The southern Gravaneset fault* (**Fig. 5.79**) is well exposed in the lower part of the escarpment at Gravaneset (**Plate 5**). The zone affected by the displacement is about **40 cm** thick and contains a cohesive flinty off-white massive fault rock that possibly originated partly from hydrothermal quartz veining. Signs of cataclasis suggests that the fault was semibrittle. The displacement on the fault is only about **50 m** in map view (**Plate 4**). The fault dips **62° NE**.

- *__Fault 11c: The northern Gravaneset fault__* (**Fig. 5.79**) is exposed at the eastern part, near the contact between conglomerates (of the Vikafjell Formation) and the Gravaneset sandstones (**Plate 4**). The fault is brittle with a **20 cm** thick cohesive flinty off-white massive fault rock. This fault rock is very similar to the one in the southern Gravaneset fault. The apparent displacement in map view has been of the order of maximum **60 m** (**Plate 4**). The fault dips **68° SSW**.

Faults __*not*__ suitable for test of displacement

The following faults are examples of faults which are *not* suitable for an evaluation of amounts of displacement:

- *__Fault 6: The Leirvågfjellet fault__* (**Fig. 5.79**), on the Leirvågfjellet plateau, is not suitable for evaluation of amount of displacement since a large area around the intersection of the fault and the Devonian/substrate contact is not exposed. In addition to this, it is not clear where the fault continues in the eastward direction from the Leirvågfjellet plateau, or in the westward direction from the Flotatjønna lakes (**Plate 1**). The eastern segment of the fault on the Leirvågfjellet plateau has been traced in the field. In the easternmost parts of the plateau, the fault is exposed in Devonian rocks, showing only vague indications of shear fractures in a cohesive rock. This fault fabric is, however, the least developed of all the fault-fabrics exposed, and it is therefore concluded that movement on the fault is negligible. The exact orientation of the fault is uncertain, but the vertical orientation shown on the eastern profile (**Plate 2**) is probably reasonably correct.

- *__Fault 8: The Breiskåra–Østre Håsteinen fault__* (**Fig. 5.79**) is not suitable for test of amount of displacement, since the area at Breiskåra, where the fault meets the Devonian/substrate contact (**Plate 1**), is not sufficiently exposed, and since the fault-trace and the contact between the Devonian rocks and the Kvangagjelet syenite becomes practically parallel here, making detection of offset impossible. In the west, at the Breiskåra area, the fault has been traced in the field. Towards the east, the position of the fault is taken from aerial photos, on which the fault has a prominent signature. At the easternmost parts, the fault cannot be traced with certainty to the Devonian/substrate contact.

Breiskåra is the name of the accessible cliff-shelf positioned high up in the southern subvertical wall of the Håsteinen mountain. The shelf, which rises gently eastwards, and which can be walked all the way on a path along the mountainside, has been eroded out along the weak zone formed by the fault. The rocks at Breiskåra have, to a variable extent been subjedted to cataclasis, and the cataclasite is possibly related to the fault movements. Exposures are, however, insufficient to give the thickness of the fault zone. Since the Devonian clasts and the Kvangagjelet syenite body contain the same rock material, it is very difficult, in the field, to separate Devonian rocks from the syenite body. The rocks are in addition heavily covered with lichen. Although the exact orientation of the fault is uncertain, the fault is most likely vertical or dipping steeply NNE. For simplicity, the fault has been drawn *vertical* on the western and central profiles (**Plate 2**). The exact direction of the (finite) displacement vector is uncertain, but since the topography/fault-trace interaction makes pure strike-slip movements impossible, the movement must have been largely dip-slip. Sense of displacement is not known. Although the *amount of displacement* on the fault is uncertain, the displacement is most likely of the same small amount as for the rest of the HDM faults. On the western profile (**Plate 2**), the simplest possibility is illustrated, i.e. the fault is shown with a negligible dip-slip component.

- *Faults 9 (a, b, c): The southern, central and northern Gravanesholten-plateau faults* (**Fig. 5.79**) located to the south of Gravanesholten are not suitable for determing amount of displacement since they cannot be traced to the Devonian/substrate contact and since internal Devonian markers are, as usual, absent. The faults have E-W to ESE-WNW directed strikes and steep to subvertical dips (**Plate 4)**. The faults are defined by shallow topographical depressions, and the fault rocks are not exposed. Sense of displaceemt is not known. In addition to the three main faults displayed on the 1:1000 map (**Plate 4**), several other sub-parallel faults, which are less prominent in terms of topographic signature, are present in the area. The southernmost of the three faults will be discussed more closely:

Fault 9a: the southern Gravanesholten-plateau fault follows almost exactly along the axial trace of the Osstrupen syncline (**Plate 4**). The fault will be considered more closely, since, in the *western* part of the plateau, the orientation of bedding has the remarkable feature of changing *abruptly* instead of gradually from one limb to the other across this fault, and since it may thus be of interest to consider whether such an abrupt change may be *caused* by the faulting.

The age relationship between the fault and the HDM-D_1-deformation is uncertain, although the fault is probably post-D_1. Some **tens of metres** further to the *east*, the orientation of bedding changes more *gradually* around the fold closure (**Plate 4**). The general expression of the fault is modest compared to the faults with prominent signatures in the HDM, and the fault is defined by a single narrow fault zone. It is thus reasonable to suggest a modest displacement for the present fault, even though the amount of displacement is uncertain.

If the abrupt change of bedding-orientation should have been produced by the fault, this would have to mean that the fault movements led to "removal" of "blocks" containing those parts of the succession that contained the bedding with the "missing" (gradually changing) bedding-orientation, thereby resulting in juxtaposition of bedding with very different orientations. Such movements or "removals" are not likely to have happened, since the fault is so modestly developed, with only a single narrow fault zone, and particularly since a more *gradual* change of bedding is present just to the east. Moreover, since the *dip* of bedding is similar across the fault, the abrupt change of bedding orientation across the fault has not been produced by rotation of fault blocks either. As was explained in Sect. 5.5.2.2, the abrupt change of bedding orientation most likely formed during the F_1-folding, and not as a result of the later faulting.

- *Fault 10: Semiductile fault to the south of Gravaneset* (**Fig. 5.79**). The fault is not suitable for testing amount of displacement since the fault cannot be traced to the Devonian/substrate contact. Fault rocks in all the other faults described above (i.e. in faults both suitable and not suitable for estimating amount of displacement), appear to be of *cataclastic* types, formed by semibrittle formation mechanisms. *Semiductile* faults (i.e. "shear zones") are apparently less common within the HDM, but the present example is well exposed and well developed, and therefore suitable to illustrate the semiductile variant of the fault rocks in the massif.

The fault is located in the near vertical road-cut present to the south of Gravaneset, about **90 m** to the south of the southern boundary of the 1:200 map. (A similar fault, present some **tens of metres** north of Fault 10, will not be discussed here, as it is less prominent). The orientation of the fault is 093/90 (**Plate 4**), i.e. vertical with an E-W trend. The amount of displacement is unknown. The shear has effected the rocks in a zone at least **60 cm** wide (**Fig. 5.81a**), although the zone of prominent semiductile shear is only **30– 40 cm** wide. The conglomeratic clasts in the shear zone have been strongly deformed to lenses during the shear (**Fig. 5.81b**). In the road cut, the intensity of the shearing increases in the southward direction across the falt rock. The southernmost **5–10 cm** has a higher content of a grey-coloured monotonous, cohesive, partly ultracataclasitic to ultramylonitic fault rock. Here, the conglomeratic clasts tend to have more elongated shapes. It is uncertain whether streching has played a role in this shaping of the clasts, as the elongated shapes appear to have been produced partly by “down-milling” during the shear. Thin-section studies of the matrix show a texture with syntectonic growth of white mica and quartz in a generally fine-grained mylonitic matrix. In the road section, indications of dextral sense of shear was observed on the road-cut rock surface dipping **45° W** (rock surface shown on Fig. 5.81b, above the hammer shaft).

The ductile component of the shear zone opens for the possibility that the zone formed in relation to the D_1-deformation in the HDM, a deformation which produced the near-by situated Gravaneset parasitic F_1-folds and axial planar S_1-cleavage (see Sect. 5.5.3). The shear zone does *not* show clear crosscutting relationships with such D_1-structures, and conclusive evidence for a syn-D_1-age for the shear zone is therefore difficult to obtain.

A **5 cm** thick zone of non-cohesive fault gouge is developed along the southern side of the most intensely sheared cohesive rock. This gouge zone is the result of later movements, at a shallow crustal depth.

5.5.7.3 FAULTS ALONG THE DEVONIAN/SUBSTRATE CONTACT

The present section deals with the faults which are frequently present *along* the Devonian/substrate *contacts*. These faults must not be confused with the other type of contact-related displacement zones, described in Sect. 5.3.3, which produced the so-called "tectonically modified primary unconformable contacts". The *faults* to be described here are recognised by generally having a stronger fault fabric than the “tectonically modified contacts”, and by a usually steep to vertical dip, a dip which means that the faults appear to be oriented *independently* of the usually shallower-dipping Devonian/substrate contact (see Sect. 5.3.3). As can be seen on the main map (**Plate 1)**, the field-traced contacts around the HDM have mostly been drawn as “*tectonically modified primary unconformable contacts*”. The only major exception from this is the contact going from the Kvangagjelet canyon to the Håsteindalen valley at the southern margin of the HDM, where the contact is drawn as "*minor fault*" (**Plate 1**). This fault is not exposed, however, and will only be briefly mentioned.

Faults, however, are also locally present *along* the “tectonically modified primary unconformity” contacts, and focus will be on *these* faults. The faults of this type have not been drawn onto the map, so as to avoid any parallel "doubling", or frequent alteration, of contact signs along the contact. The reason for choosing the “tectonically modified primary unconformity”-sign — and not the fault-sign — as the general contact-sign for the HDM-margin (**Plate 1**), is found in the occurrence frequency of contact types; the “tectonically modified primary unconformity”-fabrics generally appear to occur more frequently than the fault-fabrics, allthough the distinction between the two fabrics may at times be difficult to draw on a local scale (see Sect. 5.3.3). Moreover, the “tectonically modified primary unconformity” contact sign emphasises the important fact that the HDM, in the overall picture, rests with a primary unconformable contact on its substrate (Sect. 5.3). Exposures of fabrics indicating regular faulting may, for example, be observed localy along the contact going eastward from Stegabruna at the northern margin of the HDM.

Descriptions

When present, the faults along the “tectonically modified primary unconformity” contacts may occasionally be located at the presice Devonian/substrate contacts, but they are also often located just within the Devonian rocks or the substrate rocks. This depends on the latteral extent of the exposure in each case, and on the general sub-Devonian topography in the area. The faults appear to follow ***along*** the contact, although they may also "leave" the contact itself. The thickness of the zones varies from a **few mm** to about **40 cm**, but is typically **10–20 cm**. The orientations of the zones are variable, although the faults appear to have a tendency to dip steeply in the direction below the Devonian rocks. The displacements along the margin of the HDM have produced various types of fault rocks, from semiductile to cataclastic, and have developed modest to more intens fault fabrics and fault rocks. The fault rocks are always coherent, and the lack of incoherent fault gouge suggests, according to the criteria of Sibson (1977), that the movements took place at a crustal depth larger than about **4 km**. The fault zones usually contain fragments which range in shape from pure angular breccia-types in brittle faults, to more smoothly shaped lenses in semiductile shear zones. The fragments never appear to have been subjected to proper ductile stretching, as a brittle component always appears to have contributed to "down-milling" of the size of the fragments. The matrix is often greenish in colour due to growth of chlorite.

At one locality (UTM 0565 2944) situated at the Devonian/substrate contact **125 m** to the southeast of the ridge-crest at Novene (**Plate 1**), a breccia zone with a characteristic matrix of brick-red colour is present. This breccia resembles the breccia in the fault zone below the Dalsfjord Nappe (Sect. 2.6) on the island of Atløy, Sunnfjord (comparison by the present author), which has been dated palaeomagnetically to Late Jurassic/Early Cretaceous (Sturt & Torsvik 1986, Torsvik et al. 1992; Eide et al. 1997).

The fault from Kvangagjelet to Håsteindalen has been field-traced for a limited distance eastward from the Kvangagjelet canyon. No exposures are present, however, and the fault is only marked by a topographical depression filled with extensive Quaternary deposits. This situation also appears to continue further eastwards, where talus from the adjacent Devonian cliffs to the north covers the fault. The exact orientation of the fault is uncertain, although it appears to have a steep or vertical dip. As seen on the western and central profiles (**Plate 2**), a steep dip to the north is suggested as fault orientation.

Amount of displacement

The amount of displacement on the faults along the margins of the HDM is uncertain, since there are no structural markers along the contact. The fault zones are generally quite thin and the development of fault rocks modest. The amount of displacements are therefore assumed to be minor. The marginal faults are, in these respects, compatible to the HDM-transecting faults, where displacements have been tested and found to be small. The presence of primary unconformities along the contact may also be taken to indicate that the movements are of only minor importance, as far as the juxtaposition of Devonian on its substrate is concerned.

The amount of displacement is uncertain also for the fault from Kvangagjelet to Håsteindalen, where, in addition, the "finite" movement direction is uncertain. In the western and central profiles (**Plate 2)**, the fault is drawn as a "normal" fault (since Devonian rocks are absent on the south side of the fault), and with a negligible "dip-slip" displacement (which could have been produced by both strike-slip and dip-slip type movements). Such a limited amount of displacement is compatible to the other faults in the HDM.

5.5.7.4 SUMMARY AND DISCUSSION

Summary

Categories: The faults in the HDM have been devided into two categories; those cutting *through* the massif, *which are summarised here*, and those following along the *margins* of the massif (see description above). Since no internal structural markers are present in the massif, the Devonian/substrate contact is the only marker present that may reveal the amount of displacement on the faults. Of the two categories, only the faults crosscutting the HDM and the Devonian/substrate contact, are therefore suitable for determing the amount of displacement.

Amount of displacement: For all the faults where the displacement can be determined, the movements are found to be essentially less than **50 m** in map view. For many of the faults, the movements in map view is only **10–20 m**. The fault rocks are found to be modest in terms of thickness and intensity of deformation. By comparing fault rocks from faults with *known* displacements with fault rocks from faults with *unknown* displacements, it is found that fault rocks also of the latter faults are modestly developed, suggesting that movements have been small, also on these faults. Amount of displacements have been tested where the faults cut the Devonian/substrate contact, since this is the only structural marker in the massif. If large movements were to be postulatied, dispite the small off-sets in map view, this would imply that the finite displacement vectors would have to be almost exactly parallel to the Devonian/substrate contact. This is highly unlikely, and it is thus concluded that displacements must be small.

Faults rocks: The fault rocks generally show all types of stages from semiductile shear zones via proto- to ultracataclastic zones, and to breccia (one contact-parallel locality at Novene), and all types are cohesive. (Non-cohesive fault gouge was observed at Fault no 10 that does not cut the Devonian/substrate contact). A characteristic flinty, off-white to yellow-green cataclastic fault rock is exposed in the following faults: *Fault 1: the Straumsnes fault, Fault 4: the Vassetbrekka–"Galmannsskåra" fault, Fault 7: the Osen–Teigefjell fault, Fault 11a: the southern Gravaneset fault, and Fault 11c: the northern Gravaneset fault*. These faults demonstrate that up to **40 cm** thick fault rocks may be produced with displacements of less than **50 m** in map plane, and may be used as examples on the relationship between the fault rock development and the amount of displacement. It is possible that reactivation of the faults have taken place. This may have occurred for example in the Fault no 4 “Galmannsskåra–Vassetbrekka”: shear fractures dominate most of the fault zone, but in the central parts, a narrow zone of flinty yellow-green fault rock is present. It is possibly that the flinty zone formed at a later stage than the shear fractures, although it cannot be excluded that the flinty rock was formed due to a concentration of displacements in the central parts.

The fact that the displacements on the faults are generally so small, has the very important consequence that the faults are of no importance in the interpretation of three-dimensional architecture of the Devonian massif. This issue will be discussed in the next section (Sect. 5.5.8).

Age relationships: The age relationship between the semiductile faults and the D_1-deformation in the HDM is not clear due to lack of crosscutting relationships between the respective structures. The cataclastic faults are probably post-D_1 in age, and the brittle faults that produced coherent breccia are definitely post-D_1 in age.

The orientation of the faults: The faults in the northern fold-half (Faults no 1-2-3-4-5-6-7-9a-9b-9c-10-11a-11b-11c) strikes E-W to NW-SE, and have sub-vertical to steep south- to southwestward dips (except Fault 11a, the southern Gravanest fault, which has a dip to the northeast). The same fault orientations are seen on the faults located along the Devonian/substrate *contact* in the area between Vikarimma (Vikanipa) and Straumsnes. Even the “tectonically modified primary unconformities” in this area has a general southward dip. Also the faults in the southern fold half, along the southern margin of the Håsteinen massif, have similar strike orientations, notably Fault 8: Breiskåra–Østre Håsteinen, as well as the unnumbered Kvangagjelet–Håsteindalen fault. The dip of these faults are drawn as subvertical or nortward-dipping in the profiles (**Plate 2**), but the actual orientation is uncertain. Viewed together, all faults appear to have fairly similar strike orientations, and the dips appear to be sub-vertical to steep, towards the south to southwest.

Sense of displacement: Data on the sense of displacement of the faults are presently not available, partly due to lack of exposures. One exception from this is the Osen–Teigafjell fault. In the fault contact running along the *southern* side of the Teigafjellet Substrate Wedge and down to the lake Vassetvatnet, shear sense indicators indicate dextral shear. Several other faults have produced off-set of the Devonian/substrate contact in the map plane, but such off-sets cannot be used to reveal sense of displacement, since both strike-slip and dip-slip can produce the same sense of displacement depending on the orientation of the contact surface. However, the map picture does allow one interesting observation to be made: Of the faults that cut the Devonian substrate contact, all appear to have *dextral* movements in map view (Fault 1-2-3-4-11a-11b-11c). If all the displacements do follow such a systematic behaviour, it may suggest that they formed in a common tectonic event.

Discussion

<u>*Time of displacement*:</u> Apart from the palaeomagnetically dated displacement fabric (protocataclasite?) at Gravaneset (see Sect. 5.3.3: tectonically modified primary unconformable contacts), which gave an Triassic/Early Jurassic age (Torsvik et al. 1987), it is uncertain when the various fault movements that effected the Håsteinen massif took place. The most semiductile displacements may have been related to the Devonian D_1-deformation, during which the area was burried to a temperature of **300 +/-50°** (Sect. 5.5.6). However, such a possible connection to the D_1-deformation is difficult to document, since the shear zones have not been observed to have direct crosscutting relationships with the D_1-related structures. The semicataclastic and cataclastic breccia-producing movements must be assigned to later tectonically active periods, possibly Permian, which was found to be the age of the earliest breccia-producing movements along the Dalsfjord Fault (Sturt & Torsvik 1986; Torsvik et al 1992; Eide et al. 1997), which is situated below the Dalsfjord Nappe and the Kvamshesten Devonian massif. Such zones can also be Mesozoic, as indicated by palaeomagnetic studies in the HDM (Torsvik et al 1987). Furthermore, fault movements and joints, with or without fault rocks, may be related to Tertiary tectonics, or even possibly to the post-Weichsel glacial rebound. It is reasonable to assume that the zones have been reactivated several times at constantly shallower crustal levels.

<u>*Absense of synsedimentary faults*:</u> Synsedimentary faults have been reported to be present in the Kvamshesten massif (Osmundsen et al. 1998), and the Hornelen massif (Hartz & Andersen 1997). In Håsteinen, however, no evidence have been found to suggest that the faults are synsedimentary. Rather, the following factors suggest that the faults are not synsedimentary: *1) Displacement fabrics and fault rocks indicate considerable depth*: The rocks in the fault sones, both the genuine new fault rocks and the remnant lenses of wall rock that has been captured between fracture planes, are always cohesive. This shows that the faults formed in well lithified rocks, at a considerable depth with a fairly high temperature. *2) On-lap deposits on fault surfaces appear to be absent*: Sediments deposited with depositional onlap onto palaeo-fault surfaces have not been observed.

<u>*Normal, reverse, or strike-slip movements*:</u> In the present work, the no 7: Osen–Teigafjell Fault (displacement less than **50 m**) is suggested to be either a reverse fault or a strike-slip fault. The argument for the *reverse* movement is based on the interpretation that the fault trace — when coming from the west onto the top of Teigafjell — continues on the <u>*north*</u> side of the Teigafjellet Basement Wedge (TBW). On this side, the TBW rocks are situated on top on Devonian conglomerates, separated by the southward-dipping fault. The argument for the *dip-slip* movement is based on the interpretation that the fault — when coming from the west onto the top of Teigafjell — continues on the <u>*south*</u> side of the TBW. Dextral movements on this fault would explain the juxtaposition of the Osen inlier against the Devonian conglomerats north of the inlier, and also explain the position and exposure of the Teigefjellet Substrate Wedge. Sheared rocks in the fault that continues southeastwards from the TSW, display shear sense indicators showing dextral displacement. On the main map (**Plate 1**) and on the profiles (**Plate 2**), the fault is drawn on the north side of the Teigefjellet Substrate Wedge, and thus drawn as a reverse fault. This is done to draw attention to the possibility that the fault could be a reverse fault, although interpretations are at the present stage uncertain.

Reverse faults have been reported from other Devonian massifs in western Norway. In Kvamshesten, for example, focus has been on the Kringlefjellet reverse fault (Osmundsen 1996; Osmundsen et al. 1998; Braathen 1999), which crops out in the central part of the massif. Less prominent faults were also reported present in Kvamshesten. The Kringlefjellet fault, which is oriented with an east-west strike and a northward dip of **40–60°** (Osmundsen et al. 1998; Braathen 1999), places substrate rocks on top of Devonian rocks. The thrust movement was intepreted to be the result of the N-S contraction that also folded the Kvamshesten sediments into the major syncline. The amount of displacement was uncertain; Osmundsen et al. (1998) reported that the horizontal shortening across the entire structure could approach **1 km**, whilst Braathen (1999) estimated the movement along the fault to be in excess of **200–300 m**, but probably significantly higher.

In the Håsteinen massif, field data indicating sense of displacement on the faults has generally not been obtained. Consequently, conclusions cannot be drawn as to whether the fault movements were dominated by dip-slip normal or reverse components, or strike-slip components. In map view, the displacement of the Devonian/substrate contact indicate dextral movements, but dip-slip displacements could also have also produced such a map picture. Hence, conclusions cannot be drawn as to sense of displacement in Håsteinen.

The sense of displacement on faults in the Devonian massifs has been a matter of of interest in the litterature. Reverse displacements on E-W trending faults have been interpreted as related to the N-S contraction that folded the Devonian massifs into E-W trending synclines. Various models for the stress situation during the E-W extensional movements of the Devonian basins and the N-S contraction of the basins and the surrounding areas has been presented by Roberts (1983); Chauvet & Seranne (1994); Harz & Andersen (1997); Krabbendam & Dewey (1998); Osmundsen et al. (1998); Braathen (1999); Sturt & Braathen (2001) and Larsen (2002).

5.5.8 THE THREE DIMENSIONAL ARCHITECTURE OF THE HDM

5.5.8.1 INTRODUCTION

Purpose

Section 5.5.8 deals with the three dimensional (3-D) architecture of the HDM. To illustrate this 3-D architecture, four profiles have been constructed across the massif (**Plate 2**).

Various elements in the profiles have been established in previous chapters. These include the **(1)** the orientation of bedding in the limb areas of the Osstrupen syncline, and along the axial trace (= fold axes), as well as the different clast types in the northern and southern limbs (Sect. 5.5.2); **(2)** the inliers, that form topographic highs (Sect. 5.4); **(3)** the eastward dip of bedding along the axial trace profile and the cumulative stratigraphic thickness that may be calculated from this (Sect. 5.5.5); and **(4)** the faults, with their dip orientations and amount of displacement.

The present section (Sect. 5.5.8) deals with the profiles and particularly with the construction of the position of the Devonian/substrate contact underneath the Devonian cover. This means establishing the overall depth to, orintation of, and shape of the sub-Devonian contact, and thereby the thickness of the Devonian sediments and the angular relationship between the Devonian beds and the contact, as well as how these features varies from place to place. Information on the posistion of the contact will be obtained from data coming from the sub-aerial surface of the Devonian cover and its margin, as well as from an “unfolding” exercise performed on the Osstrupen syncline. However, the section also presents other 3-D aspects, including 3-D map features and general descriptions of the Devonian profiles.

The profiles (**Plate 2**), and the main map (**Plate 1**), form the basis for the treatment of the 3-D architecture.

Terminology

The three ~N-S directed profiles have been denoted "*eastern profile*", "*central profile*", and "*western profile*" respectively, and the E-W oriented profile has been denoted the "*axial-trace profile*" (**Plate 2**). The term "*sub-Devonian contacts*" here means the unexposed contact below the Devonian sediments, and the term "*sub-aerial contacts*" refers to the exposed contacts. The term "*sub-Devonian topography*" is used for the *present-day* "topography" of the substrate rocks below the Devonian sediments, whilst the term "*palaeo-topography*" is used for the substrate topography which existed *prior* to the folding (and faulting) of the HDM and substrate.

Organisation of the subsections

Section 5.5.8 is organised as follows: After the present introduction (Sect. 5.5.8.1), the 3-D features that are evident from the main map are presented (Sect. 5.5.8.2). This is followed by a reviw of some general features in common for all the profiles, such as practical aspects of profile orientation and drawing, as well as how the surface data are reflected i the profles (Sect. 5.5.8.3). Thereafter, local details of each profile is presented (Sevt. 5.5.8.4). This is succeeded by a brief review of the particular 3-D features, in the profiles, that has been obtained from unfolding procedures (Sect. 5.5.8.5). The next subject is the unfolding of the Osstrupen syncline, and how this provides constraints for the construction of the Devonain/substrate contact in the profiles. The subject of back-rotation of the axial trace profile is also discussed (Sect. 5.5.8.6). Finally, a summary presents the main features of the the 3-D architecture (Sect. 5.5.8.7).

A discussion of possible models for how the present orientations of Devonian bedding and sub-Devonian contacts were produced will be made in Chapter 6.

5.5.8.2 3-D FEATURES ON THE MAP

The present section (5.5.8.3) draws attention to *main map* (**Plate 1**) parametres that are significant for the 3-D architecture of the HDM. These map features have relavance to 3-D aspects of the *profiles* that will be presented later. The reason for treating the map and the profiles separately, is to draw attention to important features on each of them. The most important of the hitherto not treated features will be explained in more detail in the succeeding sections.

3-D features on the maps

Several important 3-D features in the HDM may be observed directly from the map. As will be shown later, such 3-D features will be reflected in the profiles, and it is thus worthwile to first recognise the presence of the features on the map.

The orientation of bedding on the limbs. The orientation of bedding on the limbs of the Osstrupen syncline was presented in Section 5.5.2. The limbs are very straight, with bedding on the northern limb having a constant orientation of about **070/62 SE**, and the southern limb **161/62 NE**. This means that in the profiles, *bedding must be straight towards depth also*, since any significant deflections of bedding orientation at depth would also appear at the surface. The situation in the hinge zone is different, as the dip (and strike) of bedding here changes gradually. Dip of bedding in the axial-trace profile (**Plate 2**) will reflect the fold axes along the axial trace, which have been found to have an average dip of **53°** (with variations between **50** and **58°**). The present-day orientation of bedding in the limbs determins the amount of unfolding that is required to obtain the original subhorizontal bedding- orientation that existed at the time of deposition. (Strictly speaking, *unfolding* about the **53° E** axis will make all beds dip **53° E**. The “remaining” *back-rotation* necessary to reach the subhorisontal position, would mean to remove, or reverse, the tilt that the beds originally aquired due to listric rotation during the basin formation. This is discussed later).

The primary unconformity and in situ position of the HDM. A large number of exposures of the primary unconformity have been documented all around the margin of the HDM and around the inliers (Sect. 5.3). This implies that neither the massif as a whole, nor parts of it, have been decoupled from the immediate substrate. Although minor tectonic modifications have locally been observed along the contacts, the massif as a whole is situated *in situ* on its present "floor". Unfolding (and back-rotation) of Håsteinen bedding therefore implies that the *substrate* must also be rotated <u>with the same amount</u>. As we will see later, this map-feature poses important constraints on the profile modelling.

Minor displacements on faults. The faults which cut the HDM have been described in Section 5.5.7, and it was concluded that the displacements were minor or negligible. There are no intrabasinal stratigraphic *markers* in the HDM that may give structural control on the amount and direction of displacements along the faults. The only suitable structural markers are found in the far west and in the north to northwestern part of Vikafjell, where the most prominent faults cut the marker defined by the Devonian/substrate contact as well as the contact between the Devonian sediments and the Høgdene inlier. The displacements on the faults at Gravaneset and Straumsnes are less than **50 m** in map view, and the prominent faults on Vikafjellet appear to have virtually no displacements at all, or displacements of maximum **some tens of metres** in map view, at the localities where they cut the Høgdene inlier. It is therefore concluded that the faults throughout the HDM have only minor displacements. Accordingly, in the eastern profile, for example, the faults therefore cannot be used to “restore” a simpler picture by lifting up "fault blocks" in order to "connect" bedding planes and make the bedding/contact angle smaller. Consequently, the Devonian rocks must retain their relative lateral distributions during unfolding (and backrotation).

High angle between Devonian bedding and the substrate contact. As seen on the main map (**Plate 1**), high angles between bedding and the Devonian/substrate contact are present at, for instance, Vikanipa, Teigafjell and Fjellsenden. At all these localities, the Devonian bedding rests with primary unconformities against the substrate. "Unfolding" of the dipping bedding will thus not remove the high angles, but instead create *palaeo-slopes* of the sub-Devonian surface. These localities indicate that frequently, bedding is *not* parallel to the contact surface (see below). The Novene area (name from the substrate, near westernmost HDM) is the only place in the HDM where the Devonian/substrate contact dips steeper than the bedding. This particular bedding/contact angular relationship implies that after "unfolding" of the bedding, the contact will still have a considerable basinward dip. The high angle present between Devonian bedding and the substrate contact, as seen on the map, is also reflected in the profiles (**Plate 2**). As we shall see, the angle between bedding and the Devonian/substrate contact in the profiles, depends on the orientation of the *sub-Devonian contact*, which will be established from the *unfolding* exercise to be carried out on the Osstrupen syncline.

High sub-Devonian topographic relief. The essentially primary depositional Devonian/substrate contact displays large differences in altitude (**Plate 1**). This allows for a significant "sub-Devonian topography" to be inferred directly from the main map (**Plate 1**), and several examples of this aspect of the 3-D architecture will be mentioned: (i) The contact from *Straumsnes to Mannen* climbs steadily from **0 to 350 m** above sea level (**m.a.s.l.**) with primary contacts at both the bottom and the top (the Devonian bedding shows a constant orientation, and no faults disturb the steady rise of the contact). (ii) The contact along the *southern margin of the Stigen Devonian spur* climbs from **27 to almost 450 m.a.s.l.** (iii) From *Vassetvatnet to Teigafjell*, the contact climbs from **27 to 500 m.a.s.l.**, with primary unconformities at both the bottom (the southern shore of the lake Vassetvatnet) and at the top (on the top of Teigafjell). Also here there is a constant orientation of bedding, and no disturbing faults. (iv) From the southeastern side of *Vassetvatnet and northeastwards to Steinalsfjellet* (to the east of the study area), the contact climbs from **27 to about 700 m.a.s.l.** A primary contact is present at the base at the shore of the lake Vassetvatnet, but it is not known whether a primary contact is present all the way to Steindalsfjellet. (v) At Fjellsenden, the contact climbs from **25 to 350 m.a.s.l.** from the lake of *Svardalsvatnet and up to Fjellsenden*, and a large number of exposures of the primary unconformity are present at the top. (vi) The primary contact in the small valley of *Lelisdalen* is situated at **410 m.a.s.l**. (vii) The contact from *Vikane to*

Novene climbs from **0 m to 210 m.a.s.l.**, with a primary unconformity exposed on the top and apparently no faults along the contact. (viii) Along the *southern margin of the Osen inlier*, the primary contact climbs from **10 to 380 m.a.s.l.** Together, these examples show that a significant sub-Devonian topographical relief of **300–500** (possibly **700) m** can be observed from the present map picture in the HDM. It should be noted, however, that the topography presented here will be restored in an "unfolding" process, as will be carried out later.

Sharp rise and fall of Devonian/substrate contacts. Above, it was described how the Devonian/substrate contact generally displays large differences in altitude (**Plate 1**). Another characteristic feature of the HDM is that these variations may occur within very limited areas. Such areas are present around the Devonian spurs in the east, e.g. the Vikanipa spur, the Stigen spur, the Nonsnova spur, the Strupeneset spur, and the Fjellsenden spur. Between the spurs, substrate-wedges are present; notably the Teigafjellet Substrate Wedge (TSW), and the small substrate wedge on Fjellsenden. The area to the west of the Vassetvatnet lake, and the area at Fjellsenden, display particularly good examples of this kind of intense "rise and fall" of contacts around the HDM, the result of a very rugged pre-Devonian relief. This feature is also seen in the profiles (**Plate 2**).

Thin cover of Devonian rocks, particularly in the eastern parts. Some indications of the thickness of the Devonian cover may be inferred from the main map (**Plate 1**). In the eastern areas, substrate rocks crop out in the form of wedges between the spurs of Devonian rocks, and substrate is also present as inliers. When this fact is compared with the large number of exposures of the unconformity, it appears very likely that the cover of Devonian rocks is generally thin in this area. For the western part, the situation is more uncertain when based on the map, but it is possible that the Devonian cover continues down to a depth of a **couple of hundred metres** below the sea level.

5.5.8.3 GENERAL FEATURES OF THE PROFILES

Orientation and geographical position of the profiles

The ~ N-S profiles are oriented approximately orthogonal to the axial trace of the Osstrupen syncline (**Plate 1**). The reason for this choice of orientation is as follows: because bedding is oriented **070/62 SE** in the northern limb and **161/62 NE** in the southern limb of the Osstrupen syncline, profile planes that are straight will always cross some bedding planes highly obliquely. The intersecting line between profile and bedding planes will represent the apparent dip **—** a dip which will always be shallower than true dip. To ensure that the apparent dip-angle of bedding in both limbs becomes equally "reduced" due to this oblique intersection, the profiles were positioned at right angles to the axial plane. The profiles are nevertheless oriented with a slight "clockwise" rotation from the ideal orthogonal orientation, in order to cover interesting map features (such as the spurs of Devonian rocks, the "wedges" of substrate going “into” the Devonian sediments in the east, the substrate inliers within the HDM, and the main faults). Although the slight "clockwise" rotation causes the bedding dip of the northern limb to be slightly more "reduced" than that in the south, the difference is so small that it can be ignored. The specific dip-angles shown in the profiles do indeed reflect the data, as they are "intersection lines" found from stereographically plotted intersections of the bedding planes and the profile planes. The oblique intersection generally reduces the bedding dip by about **10°** in the profiles as compared to the actual dip orientation on the map. Bedding-dips in the profiles are also generally drawn to reflect local variations where present on the map. Most faults in the profiles also have "reduced" dips, reflecting the oblique intersection of the fault planes and the profile planes. Since the profiles are parallel to each other, the dip of bedding may be compared from profile to profile.

Contact symbols used in the profiles

In the profiles, the sub-Devonian contacts towards the substrate have been given a symbol denoting "*Assumed tectonically modified primary unconformity: position uncertain*" (**Plate 2**). It is very likely that large parts of this contact has preserved a primary depositional unconformity, but since the contact is situated underneath the HDM, and tectonic modifications have been observed to be common at the exposed contacts, the above symbol has been used throughout. The sub-Devonian contact symbol also corresponds to the symbol used along the margin of the HDM on the main map (**Plate 1),** but with the important difference that the sub-Devonian contact is stippled due to the high degree of uncertainty. The contact signs are smaller and closer spaced in those parts of the profile where the position of the contact is more certain.

"Gaps" left in the profiles, between the Devonian sediments and the substrate

In all the profiles, the Devonian sediments have been drawn so that the deposits terminate before the contact against the substrate is reached. This has been done to emphasise that the precise depth to and orientation of the Devonian/substrate contact is unknown. The gap between the Devonian sediments and the substrate has been drawn with variable width, and the gap is thinnest where the position of the contact is fairly certain. As mentioned in Ch. 4, no structures are given from the substrate, due to the lack of structural markers in the Høydalsfjorden Complex.

3-D features in the profiles; directly based on data from the subaerial surface

The purpose of the following paragraphs is to draw attention to some important profile features that are *based directly on primary data* recorded from the sub-aerial surface. This serves to illustrate the difference from features that are *based on unfolding-exercises* (discussed later), and not primary data. The following data reveal important 3-D features of the profiles:

Individual beds are planar towards depth. In the profiles, individual Devonian beds generally have a constant dip from the top surface to the bottom contact. This reflects the very stable orientation of bedding in the limb areas of the Osstrupen syncline (Sect. 5.5.2).

On limb-scale, beds are mutually parallel towards depth. As seen in both the N-S profiles and the axial trace profile, large numbers of planar bedding "surfaces" are also *mutually* fairly parallel. Again this reflects the very constant orientation of strike and dip of bedding in the limb areas of the Osstrupen syncline (Sect. 5.5.2).

Faults. None of the faults in the profiles have large displacements. This is in accordance with the conclusions drawn in the section on the faults (Sect. 5.5.7). The Osen–Teigafjell fault is drawn with a reverse dip-slip displacement of the order of **50 m**. A large number of such minor faults cut the HDM, but only the most prominent ones are drawn in. The faults were described further in Section 5.5.7.

Substrate inliers. The large substrate inliers are, in the profiles shown to be sub-Devonian highs, revealed by erosion. This is in accordance with the conclusions in Section 5.4. The presence of the substrate inliers implies that the cover of Devonian deposits in the whole massif is relatively thin.

Axial plane orientation. The axial plane of the Osstrupen syncline was treated in Section 5.5.2, and the orientation was shown to be vertical. This is reflected in the N-S profiles, where bedding orientation changes abruptly across the axial plane. The axial-trace-*profile* has been drawn along this axial plane.

Position of the fold closure of the Osstrupen syncline. The position of the fold closure shown in the three N-S profiles, was established in Section 5.5.2, and follows the axial trace of the Osstrupen syncline. It should be noted that the position of the fold closure in the N-S profiles is to some degree approximate, reflecting the situation on the main map (**Plate 1**). On this map, the axial trace has been drawn, as a fairly straight line, between the two well defined fold closures located at the Gravanesholten plateau in the west and at the Osen area in central HDM, and then extended eastwards with the same trend.

Hinge zone. A narrow hinge zone, which on the main map (**Plate 1**) is seen as a zone where bedding gradually changes orientation towards the Osstrupen fold closure, is present in the northern fold-half between Straumsnes and Osen (Sect. 5.5.2). The central N-S profile displays curved bedding related to this zone. Towards the east, the hinge zone is interpreted to wedge out. In the central and eastern profiles, the hinge zone is therefore absent, as shown by the straight bedding planes at the two fold closures.

Different clast types in north and south. In the west, the boundary between the meta-psammitic clasts of the Vikafjell Formation in the north, and the meta-igneous clasts of the Blåfjell Formation in the south, is gradational in km-scale, and the two formations are thus mixed in this transitional zone. In the areas to the east of Osen, a fairly sharp boundary is present between the formations (Sect. 5.2). This may have important consequences for the 3-D development (see the following sections).

Nature of the sub-Devonian contact. In the profiles, the nature of the contact underneath the Håsteinen massif is drawn in. Indications on the nature of this contact has been obtained from the observations made along the subaerially exposed Devonian/substrate contact. Although the primary unconformity around the Håsteinen massif is exposed at numerous localities, the sub-aerial marginal contacts are usually found to be "tectonically modified primary unconformable contacts", as illustrated on the main map (**Plate 1**). In accordance with this, the same contact designation has also been used in the *profiles*.

5.5.8.4 PRESENTATION OF EACH PROFILE

Features of each profile

The following paragrafs give a presentation of the different profiles with emphasis on their general features. The profiles will be described separately in the following order: the eastern, central, western, and axial-trace profile. The general features in common for all the profiles have been described above. Note that on **Plate 2**, the axial traces of the three N-S profiles have been “vertically” alined above each other on the plate. This allow a 3-D-resembling view through the HDM from east to west by viewing first the eastern, then the central and finally the western profile.

Features in the eastern profile

The eastern profile is characterised by a relatively thin cover of Devonian rocks. The Stigen(/Leirvågfjellet) spur and the Nonsnova spur is separated by substrate outcrops, and bedding in these spurs are quite steep. This creates a strange bedding-contact geometry which theoretically could have been "solved/simplified" in an easy way if adjacent faults were, for example, used to “restore” ("lift") the Nonsnova or the Stigen(/Leirvågfjell) Devonian Spur blocks "upwards", so that bedding in these blocks could join the bedding of the Teigafjell area, and thereby restore a simple relationship between the angles of bedding-dip and contact-dip, i.e. “restoring” the Devonian bedding into a simple and continuous fold limb “body”. ***However, since the faults have only minor movements they are not responsible for the distribution of the Devonian rocks along the profile***. Hence, when executing unfolding of the Devonian layering, the spurs, for example, have to stay fixed relative to each other during the unfolding.

The contact below Nonsnova and Stigen/Leirvågfjellet spur is drawn as a relatively straight line between the outcropping sub-aerial Devonian/substrate contacts. As will be shown later, this is the optimal orientation to avoid unrealistic overhang surfaces after unfolding. As soon as the Devonian sediments are modelled as going deep into the substrate, large overhang problems arise after "unfolding". This relationship suggests that the Devonian cover forming the whole ridge of the Stigen/Leirvågfjellet spur is present only as a very thin carpet on an eastward sloping sub-Devonian surface, a surface representing a sub-Devonian continuation of the hillside presently outcropping to the south of the spur. In the profile, the Devonian sediments between Teigafjell and the lake Svardalsvatnet is also drawn to be thin. This reflects the fact that the Teigafjell substrate wedge, the Litleteigen substrate inlier and the substrate of Al-lia crop out within or just to the east of the profile, thereby suggesting a thin cover. Between Teigafjell and Fjellsenden, the contacts are partly bedding-parallel. The Devonian cover is also thin in the Fjellsenden area, and this is compatible with the presence of the substrate wedge on Fjellsenden. As will be illustrated later, "unfolding" of the Osstrupen syncline thus creates at Fjellsenden a fairly steep palaeoslope. The Devonian rocks were probably situated just above the present subaerial surface of the substrate between the Stigen spur and Nonsnova, and on the top of the Teigafjell Substrate Wedge, as is indicated by the stippled “air” line on the profile.

It has earlier been suggested that a "mountain", denoted the "sub-Devonian Teigafjell", appears to be present below the Devonian cover (Sect. 5.4.4.2), and such a sub-Devonian mountain is indicated in the profile. The sub-Devonian topography is generally very irregular, although the envelope surface is semihorizontal in km-scale.

The fault on the northern side of the Teigafjellet Substrate Wedge (TSW) is drawn with an apparent, minor reverse movement. However, as discussed in Section 5.5.7, such an apparent displacement could also have been accomplished by lateral movements.

As described in Sect. 5.5.2.3, the “frequent rising and falling” style of the Devonian/substrate contacts in the HDM (e.g. along the Stigen, Nonsnova and Strupeneset Devonian spurs, **Plate 1**) was interpreted by Torsvik et al. (1987) to indicate tight folding of the contact into several local synclines and anticlines, adding up to a larger Devonian synclinorium. However, Torsvik et al. (1987) presented no evidence in support of this hypothesis of repeated tight folding, and as explained in Sect. 5.5.2.3, the present data from the area show that this interpretation must be rejected. Thus, the profile contains no such local synclines and anticlines.

As shown on the profile, the angle between the Devonian bedding and the Devonian/substrate contact surface is quite high at the Stigen spur, Nonsnova, and Fjellsenden.

At the Stigen spur the orientation of the dip is slightly more variable than in the rest of the profile. The bedding-dip here varies between **76°** and **41° SSE**, with an average dip of **58° SSE**. In the Nonsnova spur the dip is about **53° SSE**, and to the south of the TSW, the dip is about **49° SSE**. On Fjellsenden, the dip is **55° NNE** in northern parts, and **53° NNE** in the southern part (see also **Plate 1**). The fold closure of the Osstrupen syncline is indicated in the profile, having a subvertical axial plane.

In the three N-S profiles, the opening angle of the Osstrupen syncline reflects the local dips of bedding, but since the bedding orientation is fairly constant along the axial trace, this angle will only show slight variations from profile to profile. The opening angle is **84°** in the eastern profile, and **82°** in the central profile. Since the bedding-dip illustrated in the north and south limb is "reduced" with **10°** in the N-S profiles compared to reality, due to the oblique intersection between bedding and profile planes, the apparent opening angles in the profiles are around **80°**, instead of the actual **102°** reported in Sect. 5.5.2.2. The fold closure in the western profile has a varying opening angle, due to the curved bedding in the inge zone of the Osstrupen syncline.

Features in the central profile

In the central profile (**Plate 2**), the cover of Devonian sediments is thin in the northern limb, and thicker in the south. The reasons for the thin cover at *Vikanipa*, and between the *Osen inlier* and *Vassetbrekka*, is the same as for the area between *Teigafjell* and *Leirvågfjell* in the eastern profile. This means that the contact is best drawn directly between the exposed Devonian/substrate contacts, so as to avoid overhangs after "unfolding". This contact orientation therefore probably best reflects the pre-fold situation. The presence of the

Vassetbrekka substrate wedge and the *Osen inlier*, in itself suggests that the thickness of the Devonian cover cannot be large. Note that a small Devonian spur is present to the north of the Vikanipa spur (see also **Plate 1**). Note also the apparent small reverse movement indicated on the fault at the northern margin of the Osen inlier. However, as noted in Section 5.5.7, the apparent reverse displacement might have been formed by lateral movements. In the area to the south of the lake Svardalsvatn, the thickness of the cover is larger, but the position of the sub-Devonian contact is here generally more uncertain, as indicated with question marks in the profile. Northwards towards the fold closure, the stippled line, marking the Devonian/substrate contact in the southern half, is drawn with smaller intervals between stipple points, indicating that the position of the contact is less uncertain here. The southernmost part of the central profile crosses the highest mountain peak in the HDM (Østre Håsteinen, **965 m**). At the southern margin of the Osen inlier, the unconformity is drawn parallel to bedding, based on the intersection of the contact and the topography in the valley at the central areas of the map appearance of this contact. Devonian rocks were probably situated just above the present surface of the Osen inlier, as is indicated by the stippled "air" line on the profile. The sub-Devonian topography is less pronounced in the central profile than in the eastern profile.

The angle between bedding and the substrate contact is high below the areas of Vikanipa and lake Flotatjønn, and also below the southern fold half. In Vikanipa, the dip of bedding varies between **62–63° SSE**, with **70° SSE** in the far north. Between Vassetbrekka and the Osen Inlier, bedding dips **52° SSE** in the northern part and **60° SSE** in the southern part. To the south of the Osen Inlier, the dip is **61° SSE**. In the southern fold-half, the dip is **51 ° NNE** in the south, changing gradually to **48° NNE** in the north, based on the bedding orientation to the south of Nonsnipa and at Fjellsenden. At the Neset penninsula at the southern side of the lake Svardalsvatnet, two single measurements give anomalous bedding-dips of **30** and **34° NNE**. They are considered merely as local deviations (see Sect. 5.5.2.2), and are thus not drawn into the profile. In the profile, the opening angle of the Osstrupen syncline reflects the local bedding orientations, and the axial plane has a vertical orientation.

Features in the western profile

In the western profile (**Plate 2**), the Devonian cover is drawn as fairly thick, except for the central parts. The reason for the thinner cover in the central parts, is the presence of the *Osen inlier* just to the east of the profile plane, where the contact emerges at the daylight surface. Elsewhere, the position of the sub-Devonian contact in this profile is rather uncertain, as indicated by the question marks in the contact signs, except where the *Høgdene inlier* crops out in the northern slope of *Vikafjell*.

In the southernmost part of the profile, the *Kvangagjelet syenite* is present. The interpretation of the syenite is uncertain based on the data available: it is either a landslide into the Håsteinen deposits, or an integral part of the substrate. Therefore, since the position of the *Kvangagjelet syenite* is uncertain (see Sect. 5.4.5), no Devonian/substrate contact has been drawn near the syenite. If the syenite is a Devonian landslide, and thereby an integral part of the Devonian deposits, the contact should be drawn *below* the syenite.

Alternatively, if the syenite is an integral part of the substrate, the contact might be drawn above the syenite. Of these two alternatives, it is considered slightly more likely that the syenite is a landslide.

In the profile, high apparent angles between the Devonian bedding and the Devonian/substrate contact is only present in the central part. This is because the contact surface in the profile has been drawn as a gentle syncline. Unfolding nevertheless produces a substrate palaeo-mountain, against which the Devonian sediments apparently were deposited. Contrary to the eastern and central profile, however, the western profile is drawn without a prominent local sub-Devonian topography, as no information on this subject is available.

In the northern limb, the dip of bedding in the profile is **55° SSE** at Vikafjell, changing gradually to **62° SSE** at Rindane (**Plate 2**) further south. In the southern fold-half, bedding is drawn as dipping about **48° NNE** throughout. The curved bedding in the hinge zone in the northern fold-half is indicated just north of the fold closure. The axial plane is still vertical, and the interlimb angle reflects the local orientation of bedding.

Features in the axial-trace-profile

The axial-trace profile shows the constant dip of bedding, i.e. the plunge of the fold hinge, along the axial trace of the Osstrupen syncline. In Section 5.5.5 it was demonstrated that the bedding-normal cumulative stratigraphic thickness reaches a magnitude of **5.8 km** when calculated from the shore of Høydalsfjorden in the west to the river of Storelva at Svardal in the east. As indicated in the profile, the bedding is probably fairly thin in the areas to the east of the Osen/Nonsnipa area. This is due to the presence of subaerially exposed substrate both in the Osen inlier and at Svardal, i.e. just to the north and south of the profile.

To the west of Nonsnipa (Osen), however, the depth to the contact is much more uncertain. In the profile, the Devonian/substrate contact has been drawn fairly shallow in accordance with the situation in the more eastern parts of the massif, and once more to avoid unrealistically steep slopes on rotating bedding back to a subhorisontal position (see Sect. 5.5.8.6).

The dip of bedding changes gradually from **58° ESE** at *Gravanesholten* in the west, to **50° ESE** at *Osen*; and continues within this intervall to *Storedokka* further to the east. The deviating bedding-dips of **30** and **34° NNE** at the small *Neset penninsula* is once again not drawn into the profile, since they are interpreted as merely local primary variations (see Sect. 5.5.2.2).

5.5.8.5 3-D FEATURES IN THE PROFILES, AS OBTAINED FROM UNFOLDING PROCEDURES

3-D features in the profiles; from "unfolding" procedures

The purpose of this section is to draw attention to those particular 3-D features in the profile, that cannot be found directly from sub-aerial surface data, but that must instead be constructed from "unfolding" procedures that are to be carried out in a later section (Sect. 5.5.8.5).

Angle between the Devonian bedding and the sub-Devonian contact. In the *N-S profiles*, the apparent angle between bedding and the constructed Devonian/substrate contact is mostly **20–40°**, and the bedding is only locally parallel to the contact. Along the *axial trace profile*, this angle is around **53°**, which correponds to the average fold axis for the Osstrupen syncline. Thus, the bedding generally makes a high angle towards the sub-Devonian envelope surface. This is an extraordinary situation which has important consequences for the orientation of the contact in the pre-folding stage. This high angle between the bedding and the sub-Devonian contact is discussed in a later section dealing with the unfolding of the Osstrupen syncline, which is illustrated by unfolding of the N-S trending *eastern profile* (Sect. 5.5.8.5). As mentioned above, high angles between bedding and the Devonian/substrate contact can also be observed directly from the map (**Plate 1**), notably along the margins of the HDM.

Thickness of the Devonian cover. The thickness of the Devonian cover, i.e. the *depth* to the sub-Devonian contact, is generally not very large in the profiles. Unfolding procedures have provided constraints that have made it possible to obtain information on this subject. This is explained in the following section on unfolding constraints (Sect. 5.5.8.4).

Orientation of the sub-Devonian contacts. The sub-Devonian contacts have been drawn as straight or moderately curved lines between the points of sub-aerially outcropping Devonian/substrate contacts. This has been done to satisfy requirements related to "unfolding" of the Osstrupen syncline. Further descriptions will be given in the section on the construction of the sub-Devonian contact by means of "unfolding" (Sect. 5.5.8.5).

5.5.8.6 UNFOLDING OF THE OSSTRUPEN SYNCLINE: PROVIDING CONSTRAINTS FOR THE CONSTRUCTION OF THE DEVONIAN/ SUBSTRATE CONTACT IN THE PROFILES

General

When faced with the challenge of drawing profiles across the HDM, all the factual geological data will be surface data, as illustrated on the maps. Hense, information is available on the orientation of bedding, faults, etc., and on the appearance of the substrate rocks in the inliers and around the Håsteinen massif. But there are two important pieces of information neccessary for profile construction that is *not* know in detail; and that is (1) the ***depth*** to the Devonian/substrate contact, i.e. the thickness of the Devonian sediments, and (2) the overall ***orientation and shape*** of the Devonian/substrate contact, i.e. the angle between the Devonian bedding and the contact.

Since any modelling of the 3-D architecture of the HDM implies inference of the geometries in the non-observable sub-surface, it is crucial to identify how the available data can provide constraints for this modelling. The constraints for profile construction are of two different types:

1) The first set of constraints are given by the factual geological data that have been recorded in the Håsteinen area, and is termed "*primary data constraints for unfolding*" (see below).
2) The second set of constraints appears after attempts to unfold the Osstrupen syncline and may be termed "*unfolding-constraints for profile construction*" (see below).

Since the Osstrupen syncline is a relatively simple and uniform fold structure, and since the bedding must have been originally sub-horizontal (Sect. 5.5.2) at the time of deposition, it is possible to "unfold" the structure. Furthermore, since the Devonian sediments lies in situ on the substrate, and since the minor fault displacements may be ignored, the Osstrupen syncline apparently has not experienced tectonic "disintegration" after the folding, and the unfolding may be carried out without a prior "restoration" of the fold body.

Since the folding mechanism cannot be known for sure, the unfolding to be carried out below must be considered as a *pure geometrical exersice* and not as an attempt to reconstruct the original fold history in any detail.

Four important "*primary data constraints*", which represent fundamental constraints for the 3-D modelling of the HDM, and which have been reviewed in the previous section (Sect. 5.5.8.3), can be formulated in four short sentences:

(1) Bedding in both the northern and southern limb is oriented with a largely constant strike and dip.
(2) The sediments are situated in situ with respect to the substrate, with the primary unconformitiy preserved in numerous exposures.
(3) Faulting is only minor and plays no important role in the mutual distribution and elevation of Devonian deposits and the substrate.
(4) The presence of the substrate wedges that goes "into" the Devonian in the east, and the presence of the inliers within the HDM, show that the Devonian cover in these areas is thin.

In addition to these ***primary data constraints*** for the unfolding, another condition — termed the "*unfolding-constraint*" for the construction of the sub-Devonian contact in the profiles — can now be formulated as follows:

- When the sub-Devonian contacts are to be drawn into the profiles, they must be given an orientation which allows "unfolding", i.e. allows back-rotation of the bedding to its original primary sub-horizontal position, without creating geologically unrealistic situations. Here, the term "unrealistic" will mainly mean large *overhanging* palaeo-cliffs, since situations with the less "dramatic" *steep* to *vertical* palaeo-cliffs are not completely "unrealistic", although probably rare.

This "unfolding-constraint" is very important, as it puts rigid limits to (1) how the sub-Devonian surface may be oriented, and thus also (2) how deep into the substrate the Devonian rocks can go, without creating unrealistic overhanging surfaces in the substrate after the "unfolding".

General rules for deciding the orientation of the sub-Devonian surface

If large-scale palaeo-overhang cliffs on a general basis are assumed to be geologically unlikely geomorphologies, the following general rule for the orientation of the sub-Devonian envelope contact surface may be established: Since the bedding generally dips quite steeply towards the central E-W axis of the massif, it is – when at all geometrically possible – most likely that the enveloping depositional contact surface also dips this way. (1) (See **Fig. 5.82a**, **5.82.b,** and the western profile, **Plate 2**). However, if such a basinward dip of the contact should not be possible, so that the eveloping contact surface has to dip towards the margins of the massif (see below), the dip of the sub-Devonian envelope surface should not exceed

$$90° - X°$$

where X° is the present dip of bedding (= **62° SE or 62° NE**), since a contact-dip larger than this would give a nearly vertical palaeo-cliff or possibly a palao-overhang during the "unfolding". (2) (See **Fig. 5.82a**, **5.82.b**).

Applied to the northern limb, where bedding dips more than **60° SE**, this implies that the maximum dip of the sub-Devonian envelope surface, away from the axial trace, will be

$$90° - 60° = 30°$$

i.e. towards the N-NW, and the profile becomes increasingly "realistic" the more the contact-dip approaches a horizontal orientation or even a southward dip (i.e. approaches a bedding-parallel orientation). (**Fig. 5.82.b**). The

(1) The Novene area, where the Devonian/substrate contact is steeper than the bedding, is an exception from this.

(2) It may be noted, that in the N-S profiles, the sub-Devonian envelope surfaces that dip towards the margin of the massif will appear with a somewhat shallower dip than in "reality" due to the oblique intersection between the bedding planes and the profile planes.

same reasoning applies to the southern limb, although the sub-Devonian surface there should dip maximum **30°** towards the S-SW, and the profile once again would become more "realistic" the more the surface approaches a bedding-parallel northward dip. (**Fig. 5.82.b**).

Unfolding of the eastern profile

As an example, the eastern N-S profile will be unfolded. The purpose of this unfolding is to use it as a technique to construct the Devonian/substrate contact underneath the HDM — not to discuss different models for folding of the Devonian rocks in general terms.

The unfolding will be executed by rigid rotation about an axis *orthogonal* to the eastern profile, i.e. an E-W oriented horizontal axis, and not about the average *fold axis* for the Osstrupen syncline, plunging **53 ° ESE**. The rotation axis is placed at the intersection between the axial plane (= axial trace profile) and the Devonian/substrate contact. In practice, the profile is cut into two parts along the vertcal axial plane, and each half is rotated until bedding becomes subhorisontal. The result of the unfolding is shown in **Fig. 5.83.**

In the unfolded profile, the substrate rocks form a palaeo-mountain underneath the Devonian deposits. The Devonian sediments would have to be deposited against the palaeo-mountain, to aquire their depositional contact.

The palaeo-mountain emerges because of the four "primary data constraints "; (1) in situ position of the HDM, (2) negligible displacements on faults, (3) constant bedding orientation throughout the limbs, and (4) thin cover of Devonian deposits in the east. Unfolding of the two other N-S directed profiles would give corresponding palaeo-mountains, since the Devonian/substrate contact defines an envelope surface that is sub-horizontal prior to the unfolding, and since the Devonian bedding in the profiles dips **~ 50°** towards this envelope surface prior to unfolding (**Plate 2**).

When unfolding the Håsteinen beds, it is worth noticing that the beds were probably deposited with a slight primary inclination. The bedding in Håsteinen is defined by surfaces of alluvial fans that may have had a primary inclination of up to **~ 10°** during deposition (see Sect. 5.2.1). However, because the palaeotransport directions of the sediments are largely unknown, the exact *direction* of such a possible primary bedding-dip cannot be inferred. Transport from the northern and southern margins towards the central parts of Håsteinen is, however, possible. Although it is not known whether such a primary dip of bedding would be present in the profile, the profile has nevertheless not been unfolded to a degree where the bedding appears completely horizontal (**Fig. 5.83**).

In the unfolded situation it is easy to see that the contact surface must be drawn fairly straight between the points of sub-aerially exposed Devonian/substrate contacts, to avoid "unrealistic" overhanging surfaces. The consequence of this is that the four profiles display large bedding-contact angles in most parts of their lenghts.

Geometrically, unfolding around the horizontal rotation axis is not quite correct, and it would lead to "tearing-apart" of the fold in the fold closure, i.e. along the axial plane. However, this rotation-axis has been chosen for simplicity, since "unfolding" with rotation about the **53°** axis would be more complicated, because it would implicate not only a “downward” component in the movement of the limbs and substrate, but also a "westward" component in the movement of the units, until bedding would be standing as a planar surface with a NNE-SSW strike and a dip of **53° ESE**. An E-W profile crossing *this* bedding orientation would probably resemble the present axial-trace profile, and the present-day, pre-unfold semi-horiziontal envelope surface would in the northern limb have rotated to receive a slight dip towards the N-NW, and in the southern limb a slight dip towards the S-SW.

However, regardless of whether the **horizontal** axis or **53°** axis is used during rotation, the principal relationships between the dip of bedding and the sub-Devonian contact would be essentially the same, and since the unfolding exercise does not pretend to reconstruct the exact prefold situation, rotation about a *horizontal axis* is sufficient for the purpose of getting an idea of the position and orientation of the Devonian/substrate contact.

Fold mechanisms: Rigid rotation versus vertical shear

The fold mechanism operating during the folding of the Håsteinen Massif cannot be known for sure. Nevertheless, one fold mechanism worth considering is *vertical shear*. If we use an idealized profile that in map view is oriented orthogonal to bedding strike (see profile A-B-C of **Fig. 5.84a,** c.f. map of **Fig. 5.82a**), both the northern and southern limbs will have bedding that dips **~ 60°**. If *vertcal shear* (computer generated) is performed on the profile of **Fig. 5.84a**, the shear may be allowed to proceed until the desired result is obtained: the Devonian bedding becomes horisontal (**Fig. 5.84b**). Another important result of this vertical shear will be the emergence of a palaeomountain of substrate rocks, against which the Devonian sediments were deposited.

For comparison, the idealized profile (**Fig. 5.84a**) has been subjected to folding also by rotation about an axis trending horizontally along the axial trace (**Fig. 5.84c**). This rotation axis would correspond to the one used during the above unfolding of the Eastern profile. After rigid rotation, bedding is rotated to sub-horisontal orientation, and a palaeomountain consisting of substrate rocks is produced. The rigid unfolding produces a gap above the rotation axis. The gap indicates the amount of rock that were displaced during the folding in Devonian times. Likewise, the “removal” of the substrate palaeomountain, that would have happened during the same folding in Devonian times, would imply re-location of large substrate rock volumes. In the rigid rotation model, the thickess of bedding is preserved. This is in contrast to the vertical shear model, where the bedding thickness has become doubled compared to the beds in the profile (and to the rigid rotation model). Likewise, the palaeomountain in the vertical shear model has become twice as high as in the rigid rotation model.

Back-rotation of the axial trace

The strong rotations that has produced the present-day orientation of the Devonian/substrate contact in the Håsteinen structure, can also be demonstrated with reference to the axial trace profile (**Plate 2**). The trace of bedding in this profile, i.e. the fold axis, shows a constant plunge of **53°** over a distance of over **7**

km (see also **Fig. 5.73**). Over the same distance, a primary stratigraphic or only slightly tectonically modified contact with the substrate must be assumed, as illustrated on **Plate 2**. At first sight, it might seem as though the beds were deposited against a substrate palaeoslope. However, as we shall now see, this is not possible:

Rotating this profile back to obtain an original subhorisontal bedding orientation implies co-rotation of the attached substrate, yielding an original palaeoslope of about **50°** against a mountain which is nearly **6 km** high! (**Fig. 5.85**). If we also consider the part of the Håsteinen massif located *east* of the study area (east of lake Vassetvatnet), the present-day horizontal distance will be **14 km**. When assuming that the fold axis continues at **53°** also east of Lake Vassetvatnet, back rotation of bedding to horizontal will then produce a mountain **11 km** high, against which the beds were deposited. Furthermore, it is likely that the Håsteinen sediments continued westwards from the present western margin, in which case the mountain would be even higher. It is immediately obvious that such "en block" rotation of the bedding + substrate is unrealistic, and it shows that the Håsteinen sedimets were not deposited against such a substrate palaeo-mountain. However, it serves to illustrate the geometrical problems arising from the high bedding/contact angle in the Håsteinen massif. Models for explaining such extreme conditions will be discussed in Chapter 6.

Comparison with the Hornelen Devonian Massif

The distinctive features of the Håsteinen Massif, concerning the significance of the east-dipping bedding, can be further illustrated by a comparison with the Hornelen Massif (**Fig. 5.86a** and **5.86b**). The latter massif has a continuously eastward-dipping bedding forming a cumulative stratigraphical thickness of as much as **25 km**. However, since the Hornelen strata abut on a *shear-contact* against the subjacent Nordfjord–Sogn Detachment Zone (or — in the east — against a later superimposed brittle fault), this stratigraphic thickness is not *real*, but instead a result of the syn-depositional listric rotation of the bedding during the extensional movements. A restoration of the Hornelen beds back to horisontal would thus mean reversing this process. For the Håsteinen Devonian Massif, however, the situation is completely different, since the beds cannot slide bach on top of the substrate (Høydalsfjorden Complex), due to the depositional contact towards the substrate. This unconformity means that the stratigraphical thickness of the HDM appear to be *real*. In fact, the eastward dip angle of the HDM beds along the axial trace is much steeper than the dip of the Hornelen beds, although no detachment zone is present between the HDM sediments and the substrate. Hence, the structural geometries in the Håsteinen area cannot be explained by the models hitherto published for the Devonian basins, and the situation thus requires a completely new model to explain the formation and deformation of the HDM. This is further discussed in Chapter 6.

5.5.8.7 SUMMARY OF THE 3-D ARCHITECTURE OF THE HDM

The 3-D architecture is reflected in the four profiles (**Plate 2**). The dip of bedding, as it appears in the profiles across the HDM, reflects the structural data on the main map, although the dips in the N-S profiles are drawn shallower than in reality due to the oblique profile/bedding plane intersection. The position and orientation of the Devonian/substrate contact, drawn underneath the Devonian deposits, is based on (1) constant bedding orientation in each limb; (2) in situ position of the Devonian sediments, with respect to the substrate; (3) only minor displacements on faults cutting the HDM; (4) thin cover of Devonian sediment. An unfolding procedure on a N-S oriented profile — by means of a rigid rotation about an E-W axis — has shown that the sub-Devonian bedding/substrate contact must be drawn shallow and fairly straight, to avoid overhang-surfaces after unfolding. The unfolding of the N-S profiles will create a substrate palaeo-mountain. Vertical shear is an alternative fold mechanism that can produce subhorisontal bedding around a palaeomountain. In the profiles, the open gap drawn between the Devonian and the substrate is to remind about the uncertain position of the contact. Despite the uncertain position, it is clear that large parts of the profiles contain steeply dipping bedding impinging with a high angle against the subhorizontal Devonian/substrate envelope surface. This is particularly prominent along the *axial trace profile*, where the average bedding dips **53° E**, yielding a bedding-normal stratigraphic thickness of **5.8 km**. The presence of the primary sedimentary contact below the HDM implies that the beds form a *real* stratigraphic thickness in the sense that the beds were brought adjacent to the substrate by deposition, not tectonic movements. Therefore, rigid *en block*, back-rotation of the axial trace profile to obtain subhorisontal bedding, requires co-rotation of the attached substrate as well. Rotating the Håsteinen bedding to horisontal orientation produces a palaeomountain nearly **6 km** high, with a palaeo-mountainside dipping **~ 50°** – aganst which the Devonian sediments were deposited. If the areas east of Lake Vassetvatnet are included, the palaeomountain reaches **11 km**, and higher still if the HDM is assumed to have continued westwards beyond the present western margin. It is immediately obvious that such *en block* rotation does not re-establish the original situation that prevailed during deposition of the Håsteinen sediments. A new model is therefore needed to explain the spectacular bedding-contact geometries in the HDM. The aim of this chapter was to establish the reality of these geometries, possible models will be discussed in Chapter 6.

5.5.9 SUMMARY OF THE STRUCTURAL AND METAMORPHIC DEVELOPMENT OF THE HDM

The HDM has been deformed into the massif-wide Osstrupen F_1-syncline. The average fold axis dips **53°** towards the ESE and the straight limbs are oriented **070/62 SE** and **161/62 NE**, respectively. The axial plane is subvertical. Parasitic F_1-folds are only present in the Gravaneset sandstone unit in the far west. The fold axes of the parasitic folds are essentially parallel to the fold axis of the Osstrupen syncline. In the sandstones at Gravaneset, an axial planar cleavage with WNW-ESE oriented strike and steep dip is developed. Tectonomagnetic fabrics are parallel to this tectonic cleavage. The stratigraphic thickness of the HDM has been estimated to **5.8. km** for the studied area, and **11.2 km** if the areas to the east of the lake Vassetvatnet are included. Metamorphic minerals have been observed in the cleavage in the Gravaneset sandstone unit, and it is suggested that the rocks experienced a prograde metamorphism reaching a maximum temperature of **300 +/- 50 °C**. This would correspond to anchizone/prehnite-pumpellyite facies, or the lowermost greenschist facies. The faults affecting the HDM appear to have had only minor displacements, which have been revealed by investigations of the off-sets at the places where the faults crosscut the Devonian/substrate contacts. Investigations of the 3-D architecture of the HDM has revealed that bedding in large areas stands with a high angle towards the subhorizontal Devonian/substrate envelope contact. Because the unconformity is present below the HDM, restoration of the bedding to the subhorizontal orientation that existed during deposition gives an apparently *real* stratigraphical thickness of nearly **6 km**. Since the substrate must be rotated along with the Devonian bedding, the beds would then appear to have been deposited unconformably against a steep and **6 km** high palaeomountain slope, or possibly an **11 km** high slope if the deposits east of the study area is included in the calculations. However, such a slope is obviously an unrealistic geological feature. The hitherto published models on the formation and deformation of the Devonian massifs cannot explain the geometry formed by the high angle between bedding and contact, and a new model is therefore required.

5.6 SUMMARY AND CONCLUSIONS FOR THE HÅSTEINEN DEVONIAN MASSIF

5.6.1 SUMMARY

The present section gives a summary of the the many topics in the HDM.

Rock description (cf Sect. 5.2)

Facies: The sediments in the HDM have a bimodal clast size-distribution forming two main facies: Conglomerates/breccias constitute more than **99 %** of the massif by area. Sandstones constitute less than **1%** of the massif by area, and are located in the west-southwest, west and west-northwest.

Formations/units: The conglomerates/breccias are devided into two formations. In the Vikafjell formation to the north of the "clast-type line", more than **90%** of the clasts consist of foliated meta-psammite. The Blåfjell formation to the south of the line consists of more than **90%** magmatic clasts. In addition, there is six sandstone units. These are named (1) the Slaktarholten sandstone unit, (2) the Vikaholmen sandstone unit, (3) the Gravaneset sandstone unit, (4) the Mannen sandstone unit, (5) the Galmannskåra sandstone unit, and (6) the Stegabruna sandstone unit.

Depositional processes: The conglomerates are essentially deposited by mass flows/debris flows, and the sandstones by fluvial processes.

Depositional environment: The conglomerates are deposited on the proximal and medial part of an alluvial fan, and the sandstones are probably deposited on the distal part of the fan. The fact that the sediments were deposited in an alluvial fan environment implies that the primary depositional inclination of bedding could range up to **10°**.

Primary unconformities (cf Sect. 5.3)

A large number of primary depositional unconformities are present around the margin of the HDM. This shows that the HDM is situated in situ. The unconformity cuts the F_2-folds in the substrate.

Substrate inliers (cf Sect. 5.4)

Three substrate inliers have been recognized within the HDM. These are the Høgdene meta-psammite inlier, the Osen meta-psammite inlier, and the Litleteigen "greenstone" inlier. They have all been interpreted as sub-Devonian highs, that emerged at the present daylight surface due to erosional off-strupping of the Devonian sedimensts. The Kvangagjelet syenite at the southern margin of the HDM may be intra-Devonian if it is a landslide, but since no evidence for this has been found, the syenite can alternatively an integral part of the substrate.

The Osstrupen F_1-syncline (cf Sect. 5.5.2)

The S_o-bedding in the HDM forms a massif-wide syncline termed the Ostrupen syncline. The syncline is an upright, subcylindrical, plane fold with an average fold axis oriented **115/53** (moderately plunging) and an interlimb angle of **103°** (open fold). The axial trace is fairly straight and oriented WNW-ESE. The orientation of the fold axis is very constant along the axial trace. The northern limb has an average orientation of **070/62 SE**, and the southern limb **161/62 NE**, and both limbs are remarkably straight. A narrow hinge zone is present only on the northern side of the axial plane.

The Gravaneset parasitic F_1-folds (cf Sect. 5.5.3)

First order parasitic F_1-folds are present in the Gravaneset sandstone unit. An anticline and a syncline are exposed, (and in addition an un-exposed anticline is present). Combined, the exposed folds appear to form a Z-fold (viewing E) although they are located to the north of the axial trace and therefore should have been S-folds. For the anticline and syncline respectively, the fold axes are oriented **125/55** and **129/59**, and the axial planes **136/82 NE** and **140/83 NE**. The axial planar cleavage is oriented **112/65 SW**. The folds are congruent with the fold axis of the Osstrupen syncline. The axial plane is not coinsiding completely with the axial planar cleavage.

The tectonomagnetic fabric (cf Sect. 5.5.4)

Tectonomagnetic investigations have been carried out by Torsvik et al. (1987) in the Gravaneset sandstone unit which contains the parasitic F_1-folds and the axial planar S_1-cleavage. The tectonomagnetic foliation has the same orientation as the axial plane/axial planar cleavage in the sandstone unit.

Stratigraphic thickness (cf Sect. 5.5.5)

Based on the fairly constant dip of bedding along the axial trace, the stratigraphic thickness of the HDM (in the study area) has been calculated to **5.8 km**. If the fold axes is assumed to have approximately the same orientation also in the eastern part of the HDM, the total thickness will be **11.2 km**.

M_1-metamorphism (cf Sect. 5.5.6)

The peak of metamorphism in the HDM corresponded to anchizone/prehnite-pumpellyite facies or the lowermost greenschist facies. By correlation to the Hornelen and the Kvamshesten Devonian Massifs it is reasonable to suggest a temperature of **300 +/- 50 °C**. The pressure is more uncertain, and the heat flow in the area during deposition and deformation of the HDM is not known.

Faults (cf Sect. 5.5.7)

The HDM is effected by a large number of "faults", varying from semiductile shear zones, to brittle faults and joints. Reactivations of the faults appear to have occurred at continuously more brittle conditions. Due to lack of structural markers within the HDM, the amount of displacement on the faults can only be tested for those faults which crosscut the Devonian/substrate contact at the margins or at the substrate inliers. The displacements are found to be less than **50 m**.

3-D architecture (cf Sect. 5.5.8)

The HDM beds rests with a depositional contact on the substrate, and an *en block* back-rotation of the HDM-bedding to sub-horizontal – along with the substrate – creates a palaeomountain nearly **6 km** high (or possibly **11 km** high; or more), with a palaeo-mountain-side dipping **~ 50°** – aganst which the Devonian sediments were deposited. This model obviously does not reflect the conditions during deposition of the Håsteinen sediments, but it serves to illustrate the challenging geometries in the HDM. As the published models cannot explain these extraordinary geometries, a new model is needed. This is discussed in Ch. 6.

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Chapter 6

CHAPTER 6 SUMMARY, MODELS AND CONCLUSIONS

6.1 INTRODUCTION

Chapter 6 starts with a brief summary of the main geological data of each of the three tectonostratigraphic levels in the study area (the Eikefjord Group, the Høydalsfjorden Complex, and the Håsteinen Devonian Massif), and draws some major conclusions in terms of the tectonic processes involved (Sect. 6.2). This is followed by a short account of the *chronological* order of events, where the geological history of the three tectonostratigraphic levels is integrated (Sect. 6.3). Thereafter follows a presentation and discussion of new models for the formation of the Håsteinen Devonian Massif (Sect. 6.4). This latter section is divided into five subsections. As an introduction to Sect 6.4, the "detachment" model and the "Solundian Orogeny" model is compared and discussed (Sect. 6.4.1). Subsequently, the new models for the formation of the Håsteinen Devonian Massif are discussed (Sect. 6.4.2). This is followed by a discussion of several other subjects, notably the original size of the basin (Sect. 6.4.3); development of the Nordfjord–Sogn Detachment Zone (6.4.4); and the Devonian folding that effected the HDM as well as the subjacent Høydalsfjorden Complex and Eikefjord Group (Sect. 6.4.5). Eventually, the chapter is completed by a summary of final conclusions (Sect. 6.5).

6.2 SUMMARY OF THE GEOLOGICAL HISTORY OF EACH TECTONOSTRATIGRAPHIC UNIT

Introduction

The present section gives a short review of the data presented on the Håsteinen Devonian Massif (HDM) (Sect. 6.2.1), the Høydalsfjorden Complex (HC) (Sect. 6.2.2), and the Eikefjord Group (EG) (Sect. 6.2.3). Each of the three units have their own history of geological development, which will be outlined here. In the next major section (Sect. 6.3), these three stories will be integrated.

6.2.1 EIKEFJORD GROUP

Description: The strongly mylonitic rocks of the Eikefjord Group (EG) predominantly consist of orthogneisses of the "anorthosite-Jotun-kindred" with minor presence of mylonitic metasediments. The rocks are Proterozoic of age (**1511 +/- 64 Ma**, Abdel-Monem & Bryhni 1978) and have been correlated with the Dalsfjord Suite, which is part of the Dalsfjord Nappe situated below the Kvamshesten Devonian Massif. In the study area, the earliest geological history is present in these Eikefjord orthogneisses. The earliest (pre-mylonitc) structures are collectively assigned to a $\mathbf{D_1}$-phase, which includes a Proterozoic granulite facies mineral assemblage that has been reported from EG rocks a few kilometres northeast of the study area. In addition to the Proterozoic assemblage, it cannot be excluded that this $\mathbf{D_1}$-phase also include early Caledonian fabrics, or even fabrics related to the Scandian subduction of the Western Gneiss Region (Baltic plate) below Laurentia. During the succeding $\mathbf{D_2}$-deformation, the rocks structurally became dominated by the strong penetrative mylonitic $\mathbf{S_2}$-fabric which obliterates all older structures and shows retrogressive metamorphism from the amphibolite facies and down to middle or lower greenschist facies. Shear sense indicators in the mylonite show that the shear fabrics were produced by top-to-the-west movements. An $\mathbf{L_2}$-stretcing lineation trends WNW–ESE, with shallow plunges both ways, but mainly towards WNW. The $\mathbf{S_2}$-fabric is folded by $\mathbf{F_3}$-folds with fold axes that strend WNW-ESE and plunge shallowly in both directions although mainly towards the WNW. The fold axes are parallel to the $\mathbf{L_2}$-stretching lineations. The folded fabric has a subhorisontal envelope surface on km scale. An $\mathbf{S_3}$-fabric, axial planar to the $\mathbf{F_3}$-folds, is not developed.

Interpretation: The Eikefjord Group is interpreted to be part of the Devonian Nordfjord–Sogn Detachment Zone (NSDZ), and the $\mathbf{S_2}$-mylonite fabric to be a result of the related extensional movements. The $\mathbf{F_3}$-folding of this fabric is interpreted to be a result of N-S contraction effecting the NSDZ. Although the $\mathbf{F_3}$-folds are observed to fold the $\mathbf{S_2}$-mylonites, both the $\mathbf{D_2}$- and $\mathbf{D_3}$-features are here interpreted to have been formed during the Devonian overall top-to-the-west movements affecting the Caledonian nappe pile.

The timing of movements on the NSDZ, which is developed in the Eikefjord and Lykkjebø Groups, can be inferred from geochronological and structural considerations. Devonian regional extension in western Norway was divided into a Mode-I and a Mode-II phase by Fossen (1992, 2000), further substantiated by Fossen & Dunlap (1998) and Milnes et al. (1997). The NSDZ was assignet to the Mode-II phase. The time relationship between Scandian contraction and subsequent Mode-I and Mode-II extension was established by Fossen & Dunlap (1998), who performed ^{40}Ar/^{39}Ar muscovite dating on the Caledonian decollement zone in southern Norway. They found that the Scandian top-to-the-*east* contractional movement had occurred in the interval **415–408 Ma**. This was followed by a reverse, *top-W*, extensional Mode-I movement of the entire Caledonian nappe pile on the same decollement zone, dated to have occurred in the time interval **402–394 Ma**, and interpreted to mark the onset of regional extension related to plate divergence between Baltica and Laurentia. During the Mode-I movements, the subjacent eclogitic Lower Plate constantly rose to shallower depths, causing uplift and reduction of dip-angle also of the Mode-I zone itself. Eventually, at the later stages of the Mode-I time interval of **402–394 Ma**, the zone had become sub-horizontal instead of westward-dipping, and the resulting locking of the Mode-I zone led to formation of the Mode-II/NSDZ that cut the Caledonian nappe pile. Hense, the NSDZ developed at **~394 Ma,** both in the Eikefjord area and elsewhere.

However, a mismatch (Sect. 3.5) appears to exist between the above suggested initiation of the Mode-II/NSDZ at **~394 Ma**, and the ^{40}Ar/^{39}Ar muscovite ages of **404–398 Ma** (Chauvet & Dallneyer 1992; Andersen 1998) obtained from the top-W mylonites of the Mode-II/NSDZ in the Eikefjord–Gloppen area. The muscovite retention temperature of **350 +/- 50 °C** implicates that the **404–398 Ma** Mode-II/NSDZ shear movements occurred during lower greenschist facies conditions, apparently at the same time as the Mode-I shear of **402–394 Ma** did occur along the decollement zone below the entire nappe pile. However, in addition, to the **404–398 Ma** greenschist facies shear, the NSDZ also contains an extensively developed amphibolite facies mylonitic shear fabric, indicating that movements on the NSDZ/Mode-II had started even earlier than **404–398 Ma**, while the rocks were at a larger depth.

Early movements on the NSDZ were also suggested by Johnston et al. (2007b) based on P/T analyses and radiometric dating of the Eikefjord and Lykkjebø Groups. The P/T analyses yielded prograde, peak upper amphibolite facies conditions at **13–18 kbar**, **537–618°C**, **45–60 km**. The Sm-Nd dating of garnets from this fabric gave core ages of **425.1 +/- 1.6 Ma** and **422.3 +/- 1.6 Ma**, grouped to form the interval **425–422 Ma**; and rim ages og **415.0 +/ 2.3 Ma**, **407.6 +/- 1.3 Ma** and **414.1 +/- 1.6 Ma**, grouped to form the interval **415–407 Ma**. All these ages were interpreted to date the peak upper amphibolite facies metamorphism. The error bar of **+/-1.3 Ma** on the youngest garnet rim age of **407.6 +/- 1.3 Ma** indicates that the peak amphibolite facies may have lasted until **406 Ma**. However, Johnston et al. (2007b) chose to interpret the peak metamorphism as having occurred in the interval **425–410 Ma**. This entire metamorphic history was interpreted to have occurred before the development of the NSDZ. The following onset of the NSDZ movements gave penetrative shear fabrics in the rocks. The P/T analyses of this fabric yielded a lower amphibolite to greenschist facies of **8–12 kbar**, **519–641°C**, **30–40 km**. The maximum age of the formation of the NSDZ was taken to be **410 Ma** (i.e the above age, chosen to be end of peak amphibolite metamorphism). The minimum age of shear was taken to correspond to the $^{40}Ar/^{39}Ar$ muscovite ages previously reported from the Eikefjord and Lykkejbø Group area, referred to span the interval **402–396 Ma.** The authors argued that most of the movement occurred during amphibolite facies conditions. Since the **402–396 Ma** age interval reflects greenschist facied conditions, the authors argued that the movements must have occurred prior to this age interval. Hense, using the **400 Ma** as a mean value of the last interval, Johnston et al. (2007b) suggested that the movements along the NSDZ occurred in the time interval **410–400 Ma**. (Sect. 1.5)

The above referred ages and P/T relationships might seem to open for the possibility that the Mode-II/NSDZ was developed *earlier* than the Mode-I movements, i.e. in the opposite order of what is normally assumed. However, the normal time relationship – with Mode-I first, and Mode-II next – has been independently established from extensive field work (Fossen 1992, 2000; Fossen & Dunlap 1998; Milnes et al. 1997), suggesting that this well documented time relationship has to be maintained. The cause of the mismatch is not clear. In the present thesis, the established time relationship – with Mode-I first, and Mode-II next – forms the basis for the discussions.

6.2.2 HØYDALSFJORDEN COMPLEX

Description: The Høydalsfjorden complex (HC) consists of meta-psammitic to meta-pelitic rocks – together with well bedded (**S_0**) turbiditic meta-greywackes that have been intruded by gabbro at the oceanic stage. By correlation with the Solund–Stavfjorden Ophiolite Complex, the Høydalsfjorden rocks are assumed to have originated at around **443 +/- 3 Ma** (U/Pb, Dunning & Pedersen 1988). Structurally, the rocks are dominated by a bedding-parallel **S_1**-cleavage defined by a prograde lower greenschist facies mineralogy (chlorite grade). This composite **S_0/S_1**-fabric is folded by **F_2**-folds of curved or kink/chevron type, where the fold axes trend WNW-ESE with a shallow to steep plunge in both directions. **S_2**-fabrics are present as kink bands, and locally as crenulation cleavage. The **S_2**-fabric displays only vague mineral recrystallisation, but the mineral assemblage appears to be the same as for the **S_1**-fabric. The **F_2**-folds are cut by the unconformity below the Håsteinen Devonian Massif, and the **D_1**-and **D_2**-deformation thus occurred prior to deposition of Devonian sediments. **S_3**-mylonites are present, for example in zones along the northern margin of the Høydalsfjorden Complex, showing top-to-the-west displacement. **M_3**-metamorphism in these mylonites is defined by growth of chlorte and muscovite, whereas quartz displays both cracking and recrystallisation/subgrain development, indicating deformation at the brittle/ductile transition at **~300°C**. Hence, the mylonite-producing **D_3**-deformation occurred during substantially lower temperatures than the preceding **D_2**-deformation. Folds with fold axes trending NE–SW, i.e. at high angle to the **F_2**-fold axes, have been recorded at a few instances, and these may be **F_3**-folds. **F_4**-folds are possibly present, as the **F_3**-mylonite fabric appears to have a variable dip to the north or south.

Interpretation: The rocks probably originated in an oceanic back-arc environment, related either to an ophiolite or a melange. The **D_1**- and **D_2**-deformation structures in the Høydalsfjorden Complex are interpreted as related to the top-to-the-*east* thrust emplacement of the marginal basin/outboard terrane onto the Balto-Scandian margin during the Scandian continent-continent collision, starting at around **420 Ma**. The rocks were probably emplaced onto gneissic rocks of the Western Gneiss Region, which was either of the Jostedalen Complex-type or the Fjordane Complex-type. After the **D_2**-deformation, the Caledonian orogenic contraction ceased, and crustal-scale extensional tectonics took over. This is illustrated by the **D_3**-deformation, which is related to the top-to-the-*west*, extension-related movements. Generally, such movements occurred *either* during the **Mode-I** extension in the time interval **402–394 Ma** (Fossen & Dunlap 1998), during which the entire orogenic wedge of south Norway moved westwards on the basal décollement zone – *or* during the, essentially subsequent **Mode-II** extension, during which the displacements took place on the Nordfjord–Sogn Detachment Zone (Fossen & Dunlap 1998), probably starting in the later stages of the Mode-I time interval of **402–394 Ma**. Although the **Mode-II** origin cannot be ruled out, some factors appear to indicate that the **D_3-**deformation possibly originated during the **Mode-I** extension: The **D_3**-movements in the Høydalsfjorden Complex must have seased well before deposition of the Devonian Håsteinen sediments, since the **S_3**-mylonites are cut by the unconformity below the sediments. The HC **S_3**-mylonites reflect movements that took place around **300 °C**. With a normal thermal gradient of **30 °C/km**, this implies that **~10 km** of overburden had to be removed by tectonic and/or erosional processes before the deposition of the Devonian sediments could occur, since the mylonites are truncated by the sub-Devonian unconformity. If the deposition of the Middle Devonian sediments started at, or near, the very *beginning* of Middle Devonian times, for instance at around **390 Ma**, (– the base of Middle Devonian being at **391 Ma** according to Gradstein & Ogg 1996, or at **397 Ma** according to Gradstein et al. 2004), the **S_3**-mylonites could have been formed during the preceding **Mode-I** extension at **402–394 Ma** **(**Fossen & Dunlap 1998), i.e **12–4 Ma** prior to the deposition starting at **390 Ma**. Erosion during these **12–4** million years could then have brought the **Mode-I** mylonites to the daylight surface, hence allowing Devonian sediments to be deposited unconformably on the mylonites from about **~ 390 Ma** and onwards. The shortest time span of **4 Ma** from mylonite-formation to daylight exposure is possibly insufficient when **10 km** overburden has to be removed. Erosion rates of **1 mm/year**, i.e. **1 km /Ma**, would give exhumation rates of **4 km** during these **4 Ma**, or **8 km** if the erosion rates were **2 mm/year**. If the time range available was too short for the erosion alone to be responsible for the off-stripping, tectonic processes might have contributed. Start of deposition *earlier* than at **390 Ma** would give even less time for the S_3-mylonites to become exposed.

In theory, the scenario with deposition at **~390 Ma** or later, could also imply that the S_3-mylonites could have been related to the NSDZ/**Mode-II** extension, which presumably started during the later stages of the Mode-I interval of **402–394 Ma**. However, the lower degree of metamorphism in the HC S_3-mylonites compared to the Eikefjord Group (EG) S_2-mylonites, despite the adjacent structural positions of the two units, indicates that the HC mylonites were not formed in relation to the EG mylonites, hence, suggesting that the HC S_3-mylonites formed during the **Mode-I** movements. If the Devonian sediments were deposited *later in the Middle Devonian*, more time would be available for the erosional processes to expose the mylonites, before deposition of the sediments. Moreover, if the start of deposition of Devonian sediments is viewed as *directly linked* to the formation of the **Mode-II** NSDZ at around **394 Ma**, i.e. to the formation of the EG S_2-mylonites, then the Devonian sediments would be deposited on the exposed HC S_3-mylonites around or very soon after **~ 394 Ma**. In this case, the formation of the HC S_3-mylonites would probably *have* to be related to the regional **Mode-I** movements at **402–394 Ma**, as the time would otherwise be insufficient for the erosion to expose the HC S_3-mylonites prior to the superimposed deposition of the Devonian sediments. The possible F_4-folding must have postdated formation of the S_3-mylonites, and are therefore interpreted to be related to folding of the Håsteinen Devonian Massif.

6.2.3 HÅSTEINEN DEVONIAN MASSIF (HDM)

Description: The HDM consists of continental conglomerates and very minor sandstone units. The conglomerates were deposited by mass flows in a proximal alluvial fan environment, and the deposition of sandstones occurred by fluvial processes in a more distal part of the fans. Structurally, the HDM is folded into the upright, planar Osstrupen F_1-syncline with an average fold axis having a trend/plunge of **115/53,** and a vertical axial plane trending WNW-ESE, which follows along the axial trace from the WNW end of the massif to the ESE end. The local fold axes deviate from the mean value by less than **+/- 10°** along the axial trace. The limbs are remarkably straight, with bedding on the northern limb oriented with a strike/dip around **070/62 SE** (**+/- 10°**), and on the southern limb **161/62 NE** (**+/- 10°**). A narrow hinge zone is present only on the northern "fold half", i.e. on the northern side of the axial plane. The primary depositional unconformity between the Devonian deposits and the underlying substrate rocks is exposed at a number of places around the HDM, showing that the HDM is situated *in situ* with respect to its immediate substrate, consisting of the Høydalsfjorden Complex rocks. The cumulative bedding-normal stratigraphical thickness along the axial trace is **5.8 km**. Faults crossing the massif show only negligible displacements. The combination of (i) steeply dipping bedding, (ii) *in situ* position of the Devonian deposits, and (iii) insignificant fault movements, creates a very special situation when it comes to interpreting the structural development of the Osstrupen syncline.

Interpretation: The HDM, which is assumed to be of Middle Devonian age by correlation with the Hornelen Devonian Massif, is interpreted to have been deposited in a basin that formed on the Upper Plate as a result of the post-Scandian large-scale extensional movements that followed the Early Devonian cessation of the Caledonian orogeny. The $\mathbf{F_1}$-folding of the HDM is interpreted to have occurred during the same extension. Models for this folding is discussed later (Sect. 6.4.5).

6.3 COMBINED GEOLOGICAL DEVELOPMENT OF THE STUDY AREA

6.3.1 INTRODUCTION

The present section contains a brief outline of the combined tectono-metamorphic history of the whole area. Until now the three major units of the study area have been described in a *tectonostratigraphical* order, from the Precambrian rocks of the Eikefjord Group (at the bottom, tectonostratigraphically), via the Lower Palaeozoic rocks of the Høydalsfjorden Complex (in the middle), to the sediments of the Håsteinen Devonian Massif (on top). The descriptions of structural and metamorphic developments have shown that each of these units have very distinct and different tectonometamorphic histories, and these aspects have also essentially been treated separately up to now.

In the present section (Sect. 6.3), the tectonometamorphic development will be treated chronologically, and the Devonian deformation, which in the litterature has been inferred to affect the HDM and subjacent units alike, will be treated for the whole area in common.

6.3.2 COMBINED GEOLOGICAL HISTORY

Based on the above conclusions (Sect. 6.2), the tectonometamorphic history of the Håsteinen area may be divided into two major parts; (i) orogenic top-to-the-*east* contractional Scandian events, and (ii) post-Scandian crustal-scale top-to-the-*west* extensional events. The age relationships between the different tectonic events, based on published radiometric ages, are displayed in **Table 4.1**. Structural and metamorphic rock features associated with the events are emphasized in **Table 6.1**.

Possible pre-Scandian history in the Eikefjord Group

The earliest geological history of the study area may be found in the rocks of the Eikefjord Group, which may contain remnants of a Proterozoic geological history (Bryhni et al. 1981), as well as an Early Caledonian history. Within the study area, however, the presence of such remnants has been difficult to verify, due to the later penetrative top-to-the-west mylonitisation. In the Eikefjord Group, all such possible remnants are assigned to the $\mathbf{D_1}$-event (**Table 4.1 and 6.1**), which has not been further subdivided or discussed in the present work.

The contractional Scandian history

The contractional Scandian history can be observed in the Høydalsfjorden Complex, in the form of the $\mathbf{D_1}$- and $\mathbf{D_2}$-tectonometamorphic events producing the bedding-parallel $\mathbf{S_1}$-cleavage (**Table 4.1 and 6.1**). In addition, the Eikefjord Group contains an amphibolite facies mineralogy that could theoretically be a remnant related either to the "Jotun Nappe-type" Caledonian shallow-crustal retrogression of the Proterozoic granulite facies, or to the Scandian "subduction" of Baltica/Western Gneiss Region below Laurentia. This would depend on the exact time of formation of the amphibolite facies mineral assemblage. If, in the last case, the assemblage was formed whilst the Western Gneiss Region was on "the way down" below Laurentia, it would be Scandian and assigned to the $\mathbf{D_1}$-event in the Eikefjord Group. This alternative could correspond to results obtained by Johnston et al. (2007b), who, on the basis of radiometric dating and P/T analyses suggested that the (upper) amphibolite facies occurred at **13–18 kbar**, **537–618°C**, **45–60 km**, spanning the full time interval **~424–410 Ma**. If however, the amphibolite facies was formed when the Western Gneiss Region was on the way “back up" again – associated with plate divergence between Laurentia and Baltica – it would be part of the post-Scandian extensional history. A possible example was obtained by Johnston et al. (2007b), who suggested that NSDZ shear (lower amphibolite to greenschist facies) occurred at **8–12 kbar**, **519–641°C**, **30–40 km** in the time interval **~410–400 Ma**, (although the time interval is somewhat early compared to the documented initiation of this Mode-II shear movement at the later stage of the Mode-I interval of **402–394 Ma**, see Fossen & Dunlap 1998). Between these two end members, i.e. amphibolite facies formation on the “way down” or the “way up”, an intermediate situation exists, relating to the fact that the subducted eclogitic Western Gneiss Region did in fact experience exhumation already during the last stages of the Scandian *contractional* phase, possibly allowing retrogression to amphibolite facies in this process. In such a scenario, an additional way to exhume high-grade metamorphic rocks during the Scandian *contraction*, could theoretically be by upward “extrusion” of a wedge-shaped, large-scale rock body, along the subduction zone. This would require the body to be detached either from the “top surface” of the subducted Western Gneiss Region or from the base of the overriding Laurentian craton. This kind of upward “thrusting” along a subduction zone has been suggested by Burchfiel et al. (1992) from the Himalayas. At Stadtlandet, located in the westernmost part of the Western Gneiss Region north of the Hornelen Devonian massif, the presence of ultra high pressure (UHP) and high pressure (HP) rocks has been explained as a result of syn-contractional exhumation related to thrusting and imbrication of the rocks in the sheared zone between the subducted Western Gneiss Region and the above-riding Laurentian continent (Terry & Robinson 2004).

The Devonian extensional history

General. The post-Caledonian Devonian tectonics of the study area operated in two different styles, where style 1 was the top-to-the-west *extension-related movements* that led to formation of the Devonian basins as well as to the juxtaposition of the different units, and style 2 was the N-S *contractional deformation* that led to folding of the same Devonian sediments and the substrate. The time relationship between the extensional movements and the folding is not entirely clear, but it appears reasonable to assume that the extensional movements did at least *initiate* prior to the basin formation and deposition of Middle Devonian sediments, i.e. before the folding of the sediments. During the later development, extensional movements and folding were broadly coeval.

Style 1: extensional movements. In the study area, the style 1 tectonics (top-to-the-west movements) was represented by the onset of the extensional movements that yielded mylonites with top-to-the-west shear sense in both the Høydalsfjorden Complex (HC) and the Eikefjord Group (EG). In the Høydalsfjorden Complex, which is part of the Upper Plate resting on top of the Nordfjord–Sogn Detachment Zone (/Eikefjord Group, EG), the extensional movements led to formation of the top-to-the-west HC $\mathbf{S_3}$-mylonites. In a regional context, this mylonite-producing shear could have been time equivalent with the westward movement of the entire Caledonian orogenic wedge on the subjacent decollement zone (the **Mode-I** zone of Fossen and Dunlap 1998). Folds in the HC, with NE-SW trending fold axes, could be $\mathbf{F_3}$-folds, since the folds stand at high angle to the $\mathbf{F_2}$-folds, and since the trend of the $\mathbf{F_3}$-folds would accord well with top-to-the-west (northwest) shear. In the Eikefjord Group, which is an integral part of the Nordfjord–Sogn Detachment Zone, the top-to-the-west movements produced the $\mathbf{S_2}$-mylonites and the $\mathbf{L_2}$-stretching lineations, as well as the metamorphic retrogression that has made the present rock contain minerals spanning from amphibolite facies to greenschist facies. The amphibolite facies itself could have represented only a stage in a continuous retrogression from preceding higher metamorphic grades. Johnston et al. (2007b) suggested that the amphibolite to greenschist facies shear occurred at **8–12 kbar**, **519–641°C**, **30–40 km** in the time interval **~410–400 Ma**, although this is, as mentioned, somewhat early compared to the fact that the NSDZ corresponds to the **Mode-II** zone, which formed just before, or around, **~394 Ma** (Fossen & Dunlap 1998). The HC rocks is part of the Upper Plate that apparently has moved westwards *directly* on top of the NSDZ, and it its therefore possible that the base of the HC contains mylonites that formed during these movements. However, since the HC rocks appear to constitute a down-faulted block between the Standal Fault and the Sunnar Fault, the basal contact of the HC cannot be investigated, and the presence of possible NSDZ-related mylonites cannot be confirmed. Although it cannot be completely excluded that the HC $\mathbf{S_3}$-mylonites are related to the NSDZ/EG $\mathbf{S_2}$-mylonites, this possibillity is considered unlikely, since the $\mathbf{S_3}$-mylonites are cut by the unconformity underneath the Håsteinen Devonian Massif, and since the subsequent formation of the Håsteinen and Hornelen basins themselves are apparently directly related to the movements on the NDSZ.

Deposition of Devonian ORS sediments. The extension led to the deposition of large thicknesses of Devonian Old Red Sandstone sediments in the Håsteinen basin and the other west Norwegian Devonian basins. The four basins in western Norway are seperate basins at today's erosional level, and it is considered likely that the basins were also separated during their development in Devonian times. The basins developed to thicknesses of **~ 10 km**.

Style 2: folding of the Devonian sediments. As explained above, the style 2 tectonics was the *folding* of the HDM, as well as the substrate, into roughly E-W trending folds. In the litterature, the E-W trending folds in the Devonian massifs have been explained by regional N-S contraction that effected both the Devonian sediments and the subjacent units alike (e.g. Braathen 1999; Osmundsen & Andersen 2001).

However, based on recent investigations along the northern margin of the Hornelen Devonian Massif, Larsen (2002) interpreted the folding there to be a result of intrabasinal processes, where westward movements of the basin produced shear (possibly involving transpression) along the basin-controlling marginal fault, leading to folding of the bedding. It cannot be excluded that also the Håsteinen sediments became folded as a result of (i) transpression along a lateral ramp during the extensional movements. Alternatively, the Håsteinen folding may have been caused by (ii) spaceproblems arising when the beds, during westward movements, where moved from the surface to a narrower basin bottom. Yet another possibility exists, in that the folding could be a result of (iii) a *ridge-shaped* ramp in the NSDZ, above which the Devonian sediments and the rest of the Upper Plate could have passed during their westward movements. These models are explained and discussed later (Sect. 6.4.5). In the Håsteinen basin, and the other basins, the long-lasting depositional history and large basin thicknesses (**~10 km**), suggest that the deformation of the sediments might, theoretically, have been of (i) soft sediment type, (ii) brittle type or (iii) semiductile/ductile type. However, the presence of axial planar cleavage at Gravaneset in the far west, in the Gravaneset sandstone unit, suggests that the semiductile/ductile type of deformation was the most important of the three. This is discussed later (Sect. 6.4.5).

Folding of the Høydalsfjorden Complex. In the Høydalsfjorden Complex, the WNW-ESE trend of the $\mathbf{F_2}$-fold axes, that formed prior to deposition of the above-lying Håsteinen Devonian Massif (HDM), is essentially parallel to the fold axes trend of the Osstrupen syncline of the Devonian massif. Although the Osstrupen syncline appears to be the result of a considerable N-S contraction of the HDM, conclusive evidence for the presence of $\mathbf{F_4}$-folds in the Høydalsfjorden Complex, relating to this N-S contraction, have not been observed. The apparent absence of these folds is discussed later (Sect. 6.4.5). Nevertheless, the impression of varying N- and S-ward dips of the $\mathbf{S_3}$-mylonite fabric may be taken to suggest that $\mathbf{F_4}$-folds are present. At Gravaneset, the unconformity below the Devonian Gravaneset sandstone unit is folded along with the bedding in the unit, and this shows that the Høydalsfjorden Complex has been affected by the folding of the HDM.

The parallelism between the Scandian (“pre Devonian sediment”) $\mathbf{F_2}$-folds in the Høydalsfjorden Complex, and the Post-Scandian (“syn/post Devonian sediment”) $\mathbf{F_1}$-Osstrupen syncline of the HDM, shows that the two strain systems must have been coaxial during these two events, despite the time gap between them. The lack of post-$\mathbf{F_2}$ metamorphic retrogression of the Høydalsfjorden rocks, related to N-S contraction, suggests that the N-S contraction that folded the Håsteinen deposits, had a very modest tectonometamorphic effect on the subjacent Høydalsfjorden rocks.

Folding of the Eikefjord Group. The $\mathbf{S_2}$-mylonite of the Eikefjord Group has been folded by the EG $\mathbf{F_3}$-folds, and these folds are possibly related to Devonian transpressional forces operating during the overall extensional movements, forces that produced the $\mathbf{F_1}$-fold in the Håsteinen Devonian Massif, and possibly $\mathbf{F_4}$-folds in the Høydalsfjorden Complex.

6.4 MODELS. DISCUSSIONS

6.4.1 COMPARISON OF THE “EXTENSIONAL” MODEL AND THE “SOLUNDIAN OROGENY” MODEL

The present section contains a brief comparison of the "extensional" model and the "Solundian Orogeny" model, to evaluate their potential for explaining the present data base in relation to the Devonian events in western Norway. A brief outline of the two models, as well as the background for the two points of view, was given in Chapter 2 (Sect. 2.9.2 and 2.9.3).

Any model on the geological development of western Norway must be able to explain the following major data and related processes: A mega-shear zone consisting of up to **3 km** thick mylonites (the Nordfjord–Sogn Detachment Zone, NSDZ), with top-to-the-west shear sense indicators, is present along the coast of western Norway, between Sogn and Nordfjord (Fig. 2.8). On the map, the shear zone has a sinousoidal shape, with alternating “synclinal” and “anticlinal” forms, with roughly E-W trending axial traces. In the overall picture, the shear zone separates an Upper Plate (hanging wall) situated to the west, from a Lower Plate (footwall) to the east. The shear sense indicators show that the shear zone formed due to westward movement of the Upper Plate relative to the Lower Plate. The Upper Plate contains Caledonian nappes that have experienced greenschist facies as peak metamorphism in the Caledonian orogenic cycle. The rocks of the Lower Plate, which constitute the Western Gneiss Region, were subjected to eclogite facies metamorphism in the time interval **420–400 Ma** (Carswell et al. 2003b; Hacker 2007), corresponding to the Scandian phase, after which the rocks experienced rapid uplift in Early and Middle Devonian times. [(1)] On the Upper Plate, in the areas now defined by the synclinal parts, the Devonian basins with their large amounts of Old Red Sandstone sediments were deposited, partly on top of the Upper Plate (hanging wall) which lies above the megashear zone, and partly against the megashear zone itself. The “anticlinal” areas that are located between the Solund and Kvamshesten Devonian massifs, as well as between the Kvamshesten and Håsteinen massifs, contain eclogite facies rocks that are part of the Western Gneiss Region/Lower Plate. This situation, with presence of WGR rocks between the Solund, Kvamshesten and Håsteinen massifs, differs from the situastion in the corresponding area between Håsteinen and Hornelen, an area which is instead entirely occupied by mylonites of the NSDZ. Structurally, the four Devonian deposits and the substrate have been folded about essentially E-W trending fold axes.

[(1)] A few younger eclogites, notably at **389 Ma** in the Stadlandet area (Schärer & Labrousse 2003), and four ecogites at **390–370 Ma** in the Nordøyane and Sørøyane areas (Kylander-Clark et al. 2007) have ages deviating from the “normal” time interval.

In Chapter 2, (Sect. 2.9.2 and 2.9.3), the difference between the old and new versions of the Solundian models where outlined. In essence, the old version of the model erroneously interpret the NSDZ mylonites as related to top-to-the-*east* Scandian *thrusting* (Torsvik et al. 1986) – i.e. overlooking the abundant shear sense indicators that show opposite movement – whereas the new version (Sturt & Braathen 2001) acknowledged the top-to-the-*west* extensional movements on the zone. The discussion below will deal with the new version of the Solundian model.

In both the “extensional” model and the new “Solundian” model, it is assumed that extensional movements were controlling factors for the *formation* of the Devonian basins in western Norway. The scale and significance of the extension, however, is very different for the two models, a difference which can be outlined as follows:

In the "extensional" model, the eclogitic Western Gneiss Region, defining the "Lower Plate", is juxtaposed against an "Upper Plate" displaying rocks of Scandian greenschist facies, and the two are separated by the Nordfjord–Sogn detachment zone (Norton 1987, Andersen & Jamtveit 1990; Milnes et al. 1997). The detachment zone is considered to be a fundamental crustal-scale structure that formed due to crustal extension that *postdated* the Caledonian (Scandian) contraction. The crustal-scale extensional movements imply that the Baltic and Laurentian cratons had shifted from plate collision to plate divergence (e.g. Fossen 1992, 2000; Fossen & Dunlap 1998). The N-S directed contraction in Western Norway that produced the E-W trending folds in the Devonian basins, is considered to have occurred *during* the overall extensional movements (e.g. Chauvet & Seranne 1994; Krabbendam & Dewey 1998; Osmundsen & Andersen 2001). Therefore, the "extensional" model elegantly explains the rock architecture as well as the structural styles in western Norway, including the formation of the Devonian basins.

In the "Solundian Orogeny" model, the extension is considered as a less important temporary break in the *still ongoing* contractional Caledonian orogenic cycle, where the break provides the environment necessary for the deposition of the Devonian sediements. The model has been outlined by Sturt & Braathen (2001), who presented two different scenarios designated **a)** and **b)** below, for the origin of the N-S contractional forces that folded the Devonian basins and their substrate rocks, into E-W trending folds:

a) Collision between the Baltic plate and an island arc, where the island arc was suggested to have been located near Shetland. This orogenic event was seen as having led to subduction-related volcanism that was interpreted to have produced the Devonian lavas now present on the Shetlands. The collision was considered to have occurred in Mid-Late Devonian times.

As argued in Sect. 2.9.3.2 of the present thesis, several factors make the island arc – Baltica collision model of Sturt and Braathen (2001) problematic: The suggested age for the collision (Mid-Late Devonian) would correspond to the time interval **391–354 Ma** (time-scale of Gradstein & Ogg,1996) or **398–359 Ma** (time-scale of Gradstein et al. 2004), i.e. the collision was envisaged to have happened at **~ 398 Ma** or later. However, $^{39}Ar/^{40}Ar$ datings from the base of the Jotun Nappe (Fossen & Dunlap 1998) showed that the entire Caledonian orogenic wedge experienced top-to-the-*west* movements on the basal decollement zone from **~ 402 Ma** onwards (the **Mode-I** extension of Fossen & Dunlap 1998). This conclusion of Fossen & Dunlap (1998), on the start of extension, was also referred to by Sturt & Braathen (2001), although not discussed or explicitly challenged. Hence, since crustal-scale extension due to plate divergence had started at **~ 402 Ma**, it appears unlikely that an island arc, in the Iapetus ocean (!), should, coeval with the extension (or later), collide with the Baltic craton to give the "Solundian orogeny". Further objections to the island arc – Baltica collision model may appear from the fact that the extensive eclogitisation of the Western Gneiss Region, starting at **~ 420 Ma** (e.g. Griffin et al. 1985; review in Carswell et al. 2003b) and lasting to **~400 Ma** (Carswell et al. 2003a, 2003b; Root et al. 2004; Hacker 2007) (and at Stadtlandet even to **389 Ma**, Schärer & Labrousse 2003; and at Nord-/Sør-øyane to **390–370 Ma**, Kylander-Clark et al. 2007) has been interpreted to reflect subduction of the Baltic craton (here Western Gneiss Region) below the Laurentian craton (Griffin et al. 1985, Cuthbert & Carswell 1990, Cuthbert et al 1993; Brueckner & van Roermund 2004). In such a scenario, the subduction of the Western Gneiss Region would indicate full continent–continent collision from **~ 420 Ma** onwards, implying that any island arcs, that were present outside Baltica at the time, would from this stage on, experience obduction onto the Baltic craton and further thrusting as nappes within the Caledonian orogenic wedge. This would make it impossible that an island arc, located at Shetland, could collide with Baltica at **~ 398 Ma,** or later in the **Middle or Late Devonian —** i.e **~ 20 Ma** or more, after the continent–continent collision. Such a continent–continent collision, implying complete elimination of preexisting Iapetus island arcs, has also been advocated by Hacker & Gans (2005).

b) An overall N-S compression causing sinistral transpression along the Møre–Trøndelag Fault Complex. Whereas Sturt & Braathen (2001) gave a thorough presentation of model **a)**, their model **b)** was barely mentioned, and the reader was referred to Braathen (1999) for an outline of the model. In Braathen (1999), the N-S directed orogenic contractional event, assigned to the **Late Devonian–Carboniferous**, was envisaged to be a result of the combined *southward* movement of the landmass located north of the Møre–Trøndelag Fault Complex at the time, and the relative *northward* movement of southern Norway (see Fig. 10 of Braathen 1999). These opposite directed movements were believed to have led to sinistral transpression along the zone hosting the Møre–Trøndelag Fault Complex, causing contraction of the Devonian deposits in western Norway. In Braathen (1999), this N-S contraction is envisaged to have entirely postdated the extensional movements that produced the Devonian basins of western Norway.

Also model **b)** has elements that may be questioned. The extensional processes in western Norway lasted for a long period of time, starting with westward movement of the entire orogenic wedge at **~ 402 Ma** (Fossen & Dunlap 1998), i.e. shortly before the beginning of the Middle Devonian – and probably lasting at least into the lower part of the Late Devonian, which starts at **385 Ma** (time scale of Gradstein et al. 2004) (or **370 Ma**, time scale of Gradstein & Ogg 1996). This gives extension over a time range of **~ 17 Ma** (or **~ 32 Ma**), albeit, the extension may have started earlier, and lasted longer. Also the extension-related formation of Devonian basins appears to have occurred during the Middle Devonian, but again, the duration may have been longer (e.g. Milnes et al. 1997). If the extension and ORS deposition lasted until the end of the Late Devonian, i.e. until the Devonian–Carboniferous boundary at **359 Ma** (time scale of Gradstein et al. 2004) (or **354 Ma**, time scale of Gradstein & Ogg 1996), the extensional regime would have lasted for **41 Ma** (or **46 Ma**). If the extension started earlier than Middle Devonian, the duration could be even longer. During the extensional process, the basin formation and deposition of Devonian sediments in western Norway were completed.

Likewise, the subducted eclogitic Western Gneiss Region was subjected to rapid exhumation, restoring the crust to approximately normal crustal thickness by the end of the Devonian period (Osmundsen et al. 1998; Eide et al. 1999), and exposing the eclogite-containing WGR as a core complex. During this process, the plate divergence between Laurentia and Baltica was active. In this framework, it appears problematic to introduce a new **Late Devonian–Carboniferous** orogenic phase with N-S contaction to account for the folding of the Devonian basins and their substrate. Such a folding would need to affect the whole N-S distance from the Møre–Trøndelag Fault Complex region to the Solund area, or probably even to the Bergen area, a distance of **200 km** if measured southward from e.g. the northwestern tip of the Stadtlandet peninsula. Braathen (1999) suggested that the "boundary" between the "northern" and "southern landmass" was located roughly along the Møre–Trøndelag Fault Complex (MTFC). Braathen et al (2002) extrapolated the MTFC southwestwards from the Kristiansund area to the region north of Stadtlandet, positioning the MTFC **30–40 km** north of Stadtlandet. Accordingly, the area affected by folding would span a distance of **230–240 km** between Bergen and Stadt. In the **Late Devonian–Carboniferous**, when the Caledonian crust was rapidly approaching – or had aquired – a normal crustal thickness (Osmundsen et al. 1998; Eide et al. 1999), the distribution of N-S directed deformation across such large distances would be problematic. This can be illustrated by the following reasoning: ^{39}Ar/^{40}Ar muscovite plateau ages from the NSDZ or neighbouring Western Gneiss Region (WGR) rocks south of Hornelen give ages around **400 Ma** (e.g. Andersen 1998; Chauvet & Dallmeyer 1992), corresponding to **late Early Devonian** (when using time-scale of Gradstein et al. 2004) or **mid Early Devonian** (time-scale of Gradstein & Ogg 1996). These rocks recided at the ^{39}Ar/^{40}Ar muscovite blocking temperature of **350 +/- 50 °C** at the time, a temperature corresponding to the <u>lower boundary of greenschist facies</u> (Yardly 1989). Coeval with this, the **Mode-I** extension took place at **402–394 Ma** (Fossen & Dunlap 1998), which gradually gave way to the **Mode-II**/Nordfjord–Sogn Detachment Zone, that from **~ 394 Ma** apparently controlled the formation of the Devonian basins of Western Norway. The **~ 400 Ma** ^{39}Ar/^{40}Ar muscovite plateau age of NSDZ/adjacent WGR rocks implies that after this age**,** the rocks never experienced temperatures above **350 +/- 50°C**. While Devonian sediments were deposited on the Upper Plate from **~ 394 Ma** and onwards during the Middle Devonian, the eclogitic WGR/Lower Plate was constantly exhumed (e.g. Milnes et al 1997), leading to uplift also of the above-lying NSDZ. This implies that the NSDZ would constantly occupy shallow crustal positions corresponding to temperatures lower than **350 +/- 50°C**. Eide et al (1999) performed thermal modelling of alkali feldspars that were obtained from WGR rocks at Bårdsholmen, located just south of the NSDZ south of the Kvamshesten Devonian deposit, and found that the rocks had experienced a temperature of **~ 280°C** at **350 Ma**., i.e at the Devonian-Carbonifeous boundary. It seems that the NSDZ/WGR rocks had cooled to relatively low temperatures at the Devonian-Carboniferous boundary. This appears to indicate that ductile folding of the rocks due to N-S contraction at this point in time, would be difficult to accomplish, particularly for the competent gneisses of the WGR, i.e. the Lower Plate. It might be suggested that any N-S contraction under such rheological condiditons would rather lead to thin-skin thrust tectonics, particularly within the Upper Plate.

Another aspect that may seem to argue against a N-S contraction of orogenic type, is the fact that the degree of folding, as seen in the field, does not appear to show any noticable increase northwards from Sogn to the Møre–Trøndelag Fault Complex. For example, the folding of the Devonian sediments in the Hitra–Fosen area of Møre–Trøndelag region, does not appear to be more intense than the folding of the Kvamshesten Devonian Massif in Sunnfjord in western Norway. If the inferred N-S collision between south Norway and the "northern landmass" did indeed take place along the Møre–Trøndelag zone, it would be reasonable to expect a higher degree of deformation in this area than in western Norway. The lack of a clear strain gradient northwards from the Sogn area may indicaste that the N-S contraction was not of orogenic type.

The extensional detachment mylonites of the Trøndelag region show top-to-the-southwest movements (e.g. Fig. 1 of Osmundsen et al. 2002), i.e. parallel to the MTFC – instead of movements in the N-S direction. At a first glance, this may be taken as yet another argument againt the N-S contraction model. However, this particular argument would in fact not be valid, as the N-S contraction, associated with a "northern landmass" is envisaged to have occurred *subsequent* to the extensional movements (Braathen 1999).

In the literature, it has been difficult to substantiate the existence of a N-S directed orogenic event, a fact which has also been noted by Eide et al (1999) in their statement (quote) "*At this time we lack strong evidence to support a single external tectonic event ……[that may explain the folding, etc.]*". [Text in brackets added by the present author]. Numerous papers, that deal with the various aspects of the extension that formed the Devonian basins, and also with the folding of the basins and their substrate, advocate that the N-S contraction of the basins occurred <u>during the extensional movements</u> affecting western Norway (e.g. Chauvet & Seranne 1994; Krabbendam & Dewey 1998; Osmundsen & Andersen 2001). A particularly interesting contribution in these respects was made by Larsen (2002) who interpreted the folding of the Devonian beds along the northern margin of the Hornelen Devonian basin as a result of shear along the basin-controlling northern fault margin, possibly involving transpression. Hence, the folding was seen as a result of intra-basinal processes, and not as a result of regional N-S contraction.

We now return to the comparison between the "extensional model' and the "Solundian Model". The folding of the Devonian massifs and their substrate is, as indicated above, given completely contradictory explanations in the two models.

<u>In the case of the "extensional" model</u>, it is assumed that the folding occurred <u>during</u> the overall *extensional* top-to-the-west movements. In the literature, various suggestions have been given regarding the nature of the strain systems responsible, involving for example sinistral transtension along the Møre–Trøndelag Fault Complex, a situation suggested to result in a regional constrictional strain field in western Norway (e.g. Krabbendam & Dewey 1998; Osmundsen & Andersen 2001). As an interesting variant of the "extensional" model, it could be argued that the folding of the Devonian sediments was possibly caused by transpressional forces occurring at lateral ramps in the detachment. The W-E trend of the folds are compatible with folding along such W-E oriented ramps, and folding during extension appears likely. Initially, the folds in the basins may have formed when the downward-moving beds experienced reduced N-S basin width at depth. This is discussed later in the chapter (Sect. 6.4.5).

<u>In the case of the "Solundian Orogeny" model</u>, however, the folding is considered to be a result of a new "Solundian" *contractional* phase of the Caledonian Orogeny. In Version 1 of the "Solundian model", the deformation was related to an overall top-to-the-*east* deformation, i.e. a continuation of the Scandian east-

directed trusting (or, rather, SE-directed). Remnants of this view is still present in the "Solundian model, Version 2" (Sturt & Braathen 2001), where it is suggested that the folding of the Devonian basins was caused by a collision between Baltica and an island arc near Shetland. The E-W trend of the Devonian folds was not given any explanation in the model of top-to-the-east contraction (Sturt & Braathen 2001).

In stead of invoking a separate new phase of Caledonian N-S orogenic contraction, many recent workers have supported a model where the E-W trending folds of western Norway are explained as a result of sinistral transtension along the Møre–Trøndelag zone (e.g. Krabbendam & Dewey 1998; Osmundsen & Andersen 2001; Tucker et al. 2004). Such a model was also supported by Braathen et. al. (2002).

In summary, the "Solundian model", designated "**a)**" above, that suggests E-W directed contraction and collision between Baltica and an island arc near Shetland, must be rejected, as argued above. However, in their changing of the contractional forces from E-W to N-S, which Sturt & Braathen (2001) did in the model designated "**b)**" above, their model aquired similarities with the sinistral transtension model. However, the **b)**-model also implies that the N-S contraction, leading to folding in western Norway, occurred entirely ***subsequent to*** the extensional processes that fomed the Devonian basins. As mentioned above, it is clearly most likely that the folding occurred ***during*** extension, suggesting a rejection also of the Solundian orogenic N-S contraction as a cause of folding.

In conclusion, the extensional model (i.e. "extension coeval with folding") must at the present stage be considered preferable to the "Solundian Orogeny" model ("extension first, folding next"), since the former elegantly explains (i) the geological architecture, (ii) the extensional structures and (iii) the E-W trending folds in western Norway, features which have not been satisfactorily explained by the "Solundian Orogeny" model. In addition, the various versions of the extensional model are widely accepted amongst those working with the subject. Although problems still remain when it comes to explaining the *folding* of the Devonian and its substrate, as a part of the extension, models involving (i) transpression along lateral ramps in the detachment zone, (ii) spaceproblems due to narrower width at basin bottom, and (iii) a ridge-shaped ramp, has got attractive elements. These models are discussed later (Sect. 6.4.5).

6.4.2 NEW MODELS FOR THE FORMATION OF THE HDM

General

In Sect. 5.5.8 it was explained why the hitherto published models for the formation of the Devonian massifs cannot be used to explain the HDM, and the relevant data will here be briefly reviewed. Any model applied for the HDM must be able to explain the following data from the study area:

(1) The Devonian-substrate contact is a primary (depositional) unconformity, which has locally been slightly tectonised.

(2) Bedding has constant dips around **53°ESE** from the west to the east (along the axial trace) over a lateral distance of minimum **7.2 km** – probably as much as **14.0 km** – and possibly much further still, since Devonian sediments probably once continued further westwards.

(3) The irregular Devonian/substrate contact has a sub-horizontal envelope surface (as viewed E-W along the axial plane of the Osstrupen syncline, and N-S in eastern half of the study area), and the Devonian bedding thus impinges at a high angle against the subjacent unconformity.

(4) The cumulativ bedding-normal stratigraphic thickness along the Osstrupen axial plane is **5.8 km** within the study area, probably **11.2 km** if the deposits east of the study area are included, and possibly even larger if an original westward continuation of the basin is inferred. This stratigraphic thickness is *real*, in the sense that it is without tectonically induced repetitions of bedding sequences. The continuity of the stratigraphy is apparent from the fact that the Devonian sediments rest with a primary unconformity on the substrate, and from the fact that all beds have depositional contacts towards subjacent beds.

(5) Orientation of bedding is remarkably constant in each of the two limbs of the major syncline, and the hinge zone is narrow or absent, giving a characteristic “chevron” form (acute angle) on the syncline, a form which is also found in the Hornelen Devonian Massif (Grøndalen syncline).

The hitherto applied model

The Hornelen Devonian Massif can be used to illustrate the models hitherto used to explain the relationship between the bedding and the sub-Devonian contacts of the Devonian massifs. In Hornelen, the cumulativ stratigraphic thickness of the eastward-dipping strata has been estimated to **25 km**. During basin formation, this thickness has been produced by successive rotation of the initially subhorizontally deposited layers, down along a westward-dipping, basin-controlling listric fault or detachment zone. Thus, the Hornelen sediments now rest with a tectonic shear contact against the subjacent detachment zone mylonites. This is totally different from the situation in the HDM, where the sediments rest with a depositional contact against the Upper Plate rocks (see **Fig. 5.58**).[(1)] These relationships require new models for basin development and deformation, as outlined below.

6.4.2.1 "RAMP BASIN" MODEL

Description of the ramp basin model

A new model is suggested here for the formation of the HDM. The essence of the model is that the HDM is deposited on the *Upper Plate* in a socalled “ramp depression” (in the literature also called “ramp syncline” or “ramp basin”) formed by a *ramp in the subjacent detachment zone*, where the ramp-surface may be striking N-S (or with some deviation from this), i.e. roughly orthogonal to the movement direction of the detachment zone. This ramp model is illustrated in **Fig. 6.1**, where the successive stages of the extension are shown in the form of E-W profiles. The profiles have been generated by the foreward modelling/balancing/restoration program ***“2D-Move”***. The purpose of **Fig. 6.1** is to illustrate some fundamental

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[(1)] Also in the Kvamshesten Devonian Massif, the bedding stands at an angle against the Upper Plate substrate.

geometrical relationships between the detachment shape and the bedding orientation. Details on the procedure used for ***"2D-Move"*** modelling is discussed later (Sect. 6.4.2.3).

For the Håsteinen basin, such a "flat-ramp-flat" model was first applied by Vetti (1996) and Vetti & Milnes (1997). Prior to this, Norton (1987) applied a model with very steep ramp-flat-ramp surfaces to the Solund Devonian basin, to allow for deposition in a ramp-syncline. Subsequently, Osmundsen et al. (1998) used a similar model to explain bedding gemetries in the Kvamshesten Devonian basin.

The HDM ramp basin model can be explained as follows: a westward dipping ramp is present in the detachment zone (**Fig. 6.1**), and the existence of the ramp implies that the Upper Plate consists of a *thin* eastern part and a considerably *thicker* western part. As mentioned above, the detachment zone itself can be devided into four segments, a "*lower flat*" to the west, a "*ramp*" in the middle, an "*upper flat*" further eastwards, and an "*eastern listric fault*" at the easternmost termination of the detachment, where the detachment climbes to the cut-off point (line) at the surface. When the extension starts, subaerial depressions are formed at two areas on top of the Upper Plate: (i) above the listric fault in the east, and (ii) above the ramp further west. Sediments are thus deposited at both these places. It should be noted that the altitude of the Caledonian mountains, at the time, was possibly comparable to the present-day Himalayas (Milnes et al. 1997), and with such large reliefs, the sedimentation was probably rapid.

The situation at the *listric fault*: During the westward movement of the modelled Upper Plate (**Fig. 6.1**), the sediment layers deposited above the listric fault to the east, will progressively rotate to form a typical roll-over anticline with eastward-dipping strata, and in this process, a shear contact is established against the detachment zone. At a much later stage, a large stratigraphic thickness of sediments will have accumulated, and the E-dipping bedding may stand at a considerable angle to the detachment. This particular setting would correspond to the present situation in the Hornelen Devonian Massif, but not to the HDM.

The situation at the *ramp*: It is suggested that the HDM was deposited at the *ramp-depression*, and that the following development took place (**Fig. 6.1**): when the thin, eastern part of the Upper Plate started to pass over the top-edge of the ramp, a surface depression (ramp syncline) was formed above the ramp. Here, subhorizontal beds were deposited in the ramp basin, and later transported down- and westwards on the Upper Plate. During their transport down the ramp, the beds aquired, as a geometrtical necessity, an eastward dip that was gradually increasing as the movement proceeded. Eventually, the thin Upper Plate reached the base of the ramp, and with further westward movement, the Upper Plate once again became subhorizontal. When passing the ramp–lower flat transition, the sedimentary beds had reached their full eastward dip. As the extension proceeded further, more and more beds were added to the eastward-dipping succession.

During burial, the Håsteinen sediments passed from the initial surface conditions where deformation would be of soft-sediment type, via deeper levels where cementation had produced a coherent rock that would deform by brittle faulting, to the deepest levels at **~10 km** where the deformation would be largely ductile (plastic). In the cartoon models (**Fig. 6.1**), minor deformation features like faults, etc., have not been included, since the purpose of the modelling is to reproduce the overall bedding geometries.

In more details, the requirements for any potential HDM-models, outlined in points **(1)** to **(5)** above (Sect. 6.4.2, “General”), are satisfied in the following way by the ramp model:

(1) The sediments in the depression were deposited with primary (depositional) unconformity against the Upper Plate rocks situated on the ramp-slope.

(2) + **(3)** The ramp had a moderate dip, as opposed to the sediments which were deposited with a bedding that was sub-horizontal. When the beds reached the base of the ramp, they had achieved a constant, steep eastward dip, while preserving the primary unconformity all the way.

With respect to the dip angle of the Håsteinen bedding (**Fig. 6.1**), the following should be noted: the ramp has been given a dip-angle of **35°W**, which in turn has produced Devonian bedding with a dip of **35°E**. Such a bedding-dip appears to be “incorrect”, since the recorded bedding-dip along the axial trace of the Osstrupen syncline is **53°E** (or rather ESE), i.e. a dip corresponding to the plunge of the fold axis of the syncline. In the present modelling, however, the establishment of the **53°** dip is seen as the result of two different processes: i) the rotation of bedding from the initial horisontal orientation to the dip of **35°E**, is seen as a result of the *extension*, whereas ii) the remaining **18°** steepening of bedding to achieve the recorded dip of **53°E**, is considered to have happened as a result of the N-S contraction that folded the beds into the Osstrupen syncline. The reasoning behind the use of the **35°E**-dip is outlined in the next section (Sect. 6.4.2.2).

(4) The successive down-movement of the Upper Plate on the ramp-slope will – after the Plate has passed the “ramp–lower flat” transition – lead to formation of the large thicknesses of eastward dipping bedding (**5.8 km** or **11.2 km**) on the Upper Plate. It is here important to note that generally, a ramp such as this does not need to be very high to create a large cumulative stratigraphic thicness on the Upper Plate (that is positioned above the “lower flat”). As mentioned above, the Håsteinen beds rest unconformably against the substrate, and the stratigraphy is *real* in the sense that, through the entire stratigraphy, no tectonic repetitions of beds occur, and all beds rest with depositional contacts on the bed below. However, the stratigraphy is not “real” in the sense that all strata have been situated vertically on *top* of each other *at the same time*. The bedding-normal cumulative stratigraphic thickness will therefore generally not reflect the original true *vertical* thickness of the deposit. The peak metamorphic temperature in the presently exposed HDM sediments has probably been around **300 +/-50° C**, implying that the deposits possibly experienced a "real" *vertical* burial of **~10 km** (assuming a geothermal gradient of **30°C/km**). The actual original height of the ramp is unknown. However, a basin with a vertical depth of **~10 km** would imply that the ramp itself also had a vertical height of **~10 km**.

(5) The constant bedding orientation that has been recorded on the limbs of the major Osstrupen fold, is a feature which is reasonable to expect in the ramp model, since the beds are deposited against an Upper Plate that descended with a fairly constant dip-angle down the ramp. The level of present-day erosion can be envisaged in **Fig. 6.1,** where the present-day Håsteinen and Hornelen deposits would probably constitute the lower **1–2 km** of the original **~10 km** thick sediment basins shown in the **Fig. 6.1**-profiles. To be able to aquire the high angle that is present between the bedding and the substrate in Håsteinen, the sediments could not have been deposited too high up on the slope of the down-going Upper Plate, i.e. onto the less inclined area that forms the transition zone from full ramp-slope to the flat part – as the overall angle between the bedding and the contact would then become too small. Such a small angle would be preserved during the movement of the beds/Upper Plate down the ramp. While moving down the ramp, the beds would – *close* to the Upper Plate –have a shallow-dipping or *horisontal* orientation, in contrast to the normal *eastward* dip that would exist further away from the Upper Plate. The small angle would also be preserved after movement of the Upper Plate from the ramp to the lower flat. Generally, the ramp basin model implies that the Devonian piedmont topography on the Upper Plate would be transported down the ramp and burried, and in the HDM, the recorded local relief of more than **500 m**, defined by the Devonian/substrate contact, can therefore be easily explained.

Discussion of the "ramp basin" model

From a geometrical point of view, a ramp-model can fully explain the data observed in the HDM, but it is also of interest to consider whether the model can work in practice. This may be done by comparison with *analogous sand box experiments* reported in the literature. Analogue sand box modelling of an extensional setting, where an “upper plate” of sand layers is moving on a “ramp–flat–ramp” detachment geometry (e.g. McClay 1990; McClay & Scott 1991), have demonstrated that in a ramp syncline, the rotation of bedding from the initial horizontal position to a ramp-ward dip will indeed take place (**Fig. 6.2**). If we now compare with the Håsteinen area, the “sand box situation” with E-dipping beds resting on a subhorizontal Upper Plate will correspond to the present-day E-dipping beds along the axial trace of the HDM Osstrupen syncline.

Another interesting feature of the analogue sand box experiments (e.g. McClay 1990; McClay & Scott 1991), is the *absence* of extension-related faulting in the major parts of the thin part of the “upper plate” as well as in the ramp-syncline sediments lying on top (**Fig. 6.2**). In this central part of the sand box model, significant extension-related faults have only been developed at the two “ends” of the “thin upper plate” (and onlying sediments), i.e. at those parts of the “thin upper plate” that were, at the start of extension, positioned either above the ramp, or above the listric fault. The absence of faults in the major part of the “thin upper plate” region of the sand box model shows that the transport of the “thin upper plate”, from its original position on the upper flat, and down the ramp-slope, to its position on the lower flat, occurred without any noticable break-up of the “upper plate” or onlying sediments. This is in agreement with the situation observed in the Håsteinen Devonian Massif (HDM), where there is an absence of faults with “ramp-parallel trends”, i.e. absence of faults with N-S trends or trends devating somewhat from this. The fact that severe faulting did *not* effect the main part of the sand box “thin upper plate”, during its transport down the ramp from the upper flat to the lower flat, is worth noting, as it appears to be contra-intuitive.

Although significant faults related to the extensional movements on the Nordfjord–Sogn Detachment Zone appear to be absent in Håsteinen, the presence of such faults cannot be completely excluded, and the conditions for fault formation is worth considering. Since the original depth of the Håsteinen basin – measured from the basin surface to the Upper Plate (on the lower flat) – presumably was in the range of **~10 km**, it is likely that any extension-related faulting in the presently exposed Håsteinen sediments would be ductile (plastic) at these depths. At earlier stages, before the basin reached its full depth, the deformation was of brittle type at intermediate depths, and of soft sediment type at/near the surface. It is also not known whether the potential faulting occurred before, during or after the folding of the HDM, but this would not be critical. A large number of faults are present in the HDM, but they have not affected the bedding orientation, which is very constant within each limb of thw Osstrupen syncline. Most of the prominent faults have trends varying around WNW-ESE. Since structural markers are absent within the monotonous HDM conglomerate beds, the amount of displacement on these faults is very difficult to decide within the massif itself. However, many of the faults cut the Devonian/substrate contact that is present along the margin of the massif, and off-sets are here only minor, suggesting that the faults had only negligible displacements. Anyway, the WNW-ESE trend of these faults makes them poor candidates for extension-related faults, since such faults would presumably have a more N-S to NE-SW trend. If we imagine a scenario where faulting occurred prior to consolidation, the very coarse conglomerates would deform by granular flow and leave little or no sign of the faulting. Generally, the HDM does not appear to contain prominent faults/shear zones that have N-S trends, possibly with exception of the N-S directed deep valley hosting the large lake Vassetvatnet. However, since the original strike orientation of the frontal ramp that influenced the Håsteinen area is unknown, it cannot be excluded that some of the faults in the HDM be accomodation structures.

The ramp model opens for the possibility that some degree of sediment transport into the ramp-depression has been directed from west to east, i.e. opposite to the main transport direction in the nearby situated Hornelen Devonian Massif. However, since the main portion of the sediments will have depo-centra at the eastern side of the basin, this is not an objection to the model. One problem with the ramp basin model may be that new relief to support generation of sediments will not be expected to form on those parts of the Upper Plate that are situated just east of the ramp, since the Upper Plate here does not experience uplift in the same way as the Lower Plate east of the detachment cut-off line, which would be supplying a basin of "Hornelen type" in the east. However, at the upper ramp-flat edge, the successive down-movement of the Upper Plate will nevertheless generate relief, and lack of "Upper Plate" source area will therefore not occur before the entire Upper Plate has passed down the ramp. At this advanced stage, the sediments in the eastern "Hornelen-type" of basin would become resedimented (**Fig. 6.1**). In summary, it appears that the ramp basin model will work in practice.

It is also of interest to consider to what degree the ramp model is compatible with the regional geological picture in Sunnfjord. The Standalen segment of the NSDZ is located along the southern margin of the HDM. The mylonite fabric in the zone essentially dips towards the north, and shear sense indicators show "top-to-the-west" movement (Andersen & Jamtveit 1990; Krabbendam & Dewey 1998). To the north of the HDM, i.e. north of the Sunnar Fault, the NSDZ has a subhorizonal or shallowly southward dipping envelope surface, and also here the displacement has been top-to-the-west (Sect. 3.4). The following facts must form the basis for a discussion of the compatibility of the model, to the geology in Sunnfjord: **(1)** The possible ramp must have been located to the east of the HDM, since the whole Håsteinen massif (including the deposits east of the study area) is situated on Upper Plate rocks. **(2)** The eastern *termination* of the Upper Plate, i.e. the cut-off line, must have been situated to the east of the HDM, but to the west of the Hornelen Devonian Massif (Hornelen DM). This means that the possible ramp structure to the east of the HDM must either die out towards the Hornelen

DM, or turn into a lateral ramp between the HDM and the Hornelen DM. Since the envelope surface of the mylonite between Håsteinen and Hornelen appears to be sub-horizontal (Wilks & Cuthbert 1994), the ramp most likely dies out towards the Hornelen DM. The strike orientation of the possible ramp is not known, and neither is its possible original position east of the present-day Upper Plate. The HDM and the rest of the present-day Upper Plate, situated between the Standalen segment of the NSDZ and the Sunnar Fault, is believed to lie on top of the mylonitic NSDZ. Hence, the detachment mylonites form a continuous package reaching from the Standalen segment in the south, to the Hornelen DM in the north. Such a sub-horizontal orientation of the envelope surface of the mylonite fabric in this area suggests that the ramp must have been located to the east of this area. Anyway, regardless of the original position of the ramp, the present level of erosion in the Håsteinen region and the areas further east, will be so low that the ramp would probably not be present at todays surface. Generally in western Norway, the Nordfjord–Sogn detachment zone appears to reach its eastern termination at variable distances to the east of the Devonian massifs. In addition, it is reasonable to assume that the detachment zone has continued even further to the east, i.e above the high-grade rocks of the Lower Plate (Western Gneiss Region). Combined, this would provide the necessary space for the suggested ramp structure to the east of the HDM. The Standalen segment of the NSDZ, with its steep northward dip (Bryhni & Lutro 1991a, 1991b, 2000a, 2000b,; Andersen & Jamtveit 1990; Krabbendam & Dewey 1998) might have acted as a lateral ramp. A steeply north-dipping *brittle* fault, named the Standalen Fault, follows along and cut the mylonite fabric of the Standalen segment of the NSDZ. In the western parts of this brittle fault, at a position between the lake Standalsvatnet and the sea of Førdefjorden, the fault-trend turns from E-W to ENE-WSW, thus cutting the E-W directed stretching lineations and other linear fabrics in the mylonite, at a small angle. The E-W trending stretching lineations within the mylonite are parallel to the E-W trending strike direction of the northward-dipping mylonite fabric itself, which would be compatible with movements along a lateral ramp.

In conclusion, a ramp syncline model offers an elegant explanation for the complex bedding/contact geometries in the HDM. Such a ramp structure is compatible with the overall extensional regime that was present during the formation of the HDM and the other Devonian massifs in western Norway.

6.4.2.2 ESTIMATE OF BEDDING-DIP IN HÅSTEINEN PRIOR TO FOLDING

In the ”2D-Move” model of the HDM, bedding is drawn with an eastward dip of **35°** (**Fig. 6.1**). As mentioned in the previous section (Sect. 6.4.2.1), this is **18°** less than the average bedding-dip presently found along the HDM axial trend, which is **53°** (i.e. the plunge of the Osstrupen syncline, Sect. 5.5.2.2). The **35°** angle represents an estimate of the bedding dip that may have existed in a *”not folded”* HDM, i.e. prior to folding of the HDM. As shown below, the **35°** angle is obatined by calculations based on a comparison with the Grøndalen syncline of the Hornelen Devonian basin (Sect. 5.5.2.6). Such a comparison is justified by the fact that the Grøndalen syncline closely resembles the Osstrupen syncline in terms of major fold parameters such as wave-lenght, limb-dips, etc (e.g. compare **Fig. 5.50** and **Fig. 5.55**).

The **35°** angle is here used as a best approximation to the eastward bedding-dip produced by the bedding-rotation generated by the basin-controlling ramp (Sect. 6.4.2.1). The folding of the HDM, that occurred during or after extension, has steepened the bedding-dip along the axial surface from the preceding estimated **35 °,** to the present **53°**.

The calculation was carried out as follows:

A β-plot of bedding-orientations of the limbs of the Grøndalen syncline (**Fig.6.4**), yields a fold axis with a trend/plunge of **087/38**. The two large-circles of the β-plot, representing the strike/dip of bedding of the north and south limbs, were constructed in the following way: strike values were taken from **Fig. 5.55**, (which is based on the preliminary geological 1:50.000 maps NAUSTDAL (Bryhni & Lutro 1989, 1991b, 2000b). The strike values are not given on these two maps, and are thus not shown on **Fig. 5.55**. However, these <u>strike values</u> have been obtained by carefully measuring the orientation angles of the strike/dip signs, directly on the maps. <u>Dip values</u> were obtained from field work carried out by the present author. In more detail, strike values were taken from the map area close to the hinge line **(Fig. 5.55)**. For the northern limb the seven strike values are (from west to east) **053°**, **067°**, **062°**, **065°**, **058°**, **056°**, and **059°**, giving an <u>avarage strike of **060°**</u>. For the southern limb the five strike values are (again from west to east) **093°**, **124°**, **136°**, **132°**, and **153°**, yielding an <u>average strike of **128°**</u>**.** <u>The dip value of **60°** for the northern limb</u> and <u>**50°** for the southern limb</u> were obtained in the field by using the sighting method. The sighting was performed across lake Grøndalsvatnet from suitable locations on the western side of the lake (**Fig. 5.55**), looking firstly towards the east-northeast and secondly towards the southeast, to cover the northern and southern limb respectively. The dip-values obtained by the sighting method are more representative and accurate than single bedding measurements on the outcrop. The dip-values obtained at Grøndalen have thus been recorded in an area located slightly west of the area where the strike values are taken from. However, since the Grøndalen syncline along its whole length becomes tighter towards the east, the dip-values will probably be even steeper eastward from Grøndalen. Hence, the dip-values recorded from the Grøndalen area are conservative values. The values selected for strike and dip thus give a <u>northern limb</u> with a strike/dip of **060/60 SE**, and a <u>southern limb</u> with a strike/dip of **128/50 NE**.

The orientations of the southern and northern limbs yield a Grøndalen syncline fold axis with a trend/plunge of **087/38** (**Fig. 6.4**). A trend/plunge close to this was also reported by Braathen (1999), who obtained **098/39** for the fold axis around lake Kaldevatnet (**Fig. 5.55**), which is located on the axial trace of the Grøndalen syncline at a position **~ 2 km** from the eastern marginal boundary of the Hornelen deposits.

In the central region of the Hornelen Devonian Massif, i.e. in the large area located between the folded zones present along the northern and southern margins (**Fig.5.55**) of the massif, the bedding generally dips **25°** towards east (Kolderup 1926; Steel et al. 1985). This means that during the folding that produced the Grøndalen syncline, the bedding-dip was steepened from this general **25°** dip, to the **38°** dip defined by the plunge of the Grøndalen fold axis, referred above. This represents an increase of **13°**, corresponding to **52 %**.

With its plunge of **53°**, the fold axis of the HDM Osstrupen syncline appears to be the steepest of all fold axes in the Norwegian Devonian deposits, and it is reasonable to assume that this plunge is a result of a steepening process that affected bedding during the contraction that formed the syncline, a steepening process similar to the one in the Grøndalen syncline. An estimate of the bedding-dip that was present in a ”*not folded*” HDM may thus be obtained by comparison with the Grøndalen syncline: if we assume that the steepening in the HDM was **52 %**, i.e. similar to Hornelen, and we set X to be the unsteepened dip, the expression

$$X \cdot 1.52 = 53°$$

yields an unsteepened bedding-dip of

$$X = 35°$$

for the HDM. <u>This dip value is used for the “not folded” beds in the 2D-Move modelling of the HDM</u>.

As mentioned, the above calculations are based on a general ”not folded”-bedding-dip value of **25 °** for the Hornelen basin. However, since the bedding-dips of Hornelen have been reported to vary within the interval **20–25°**, it is of interest to see how sensitive the calculated ”not folded”-dip of <u>Håsteinen</u> is to the selected dip value for <u>Hornelen</u>. The effect of using **20°** instead of **25°** (reduction of **5°**) for the Hornelen bedding-dip has been calculated, coming out at **28°** instead of **35°** for Håsteinen **(Table 6.2)**, i.e. a reduction of **7 °**. In percentage, the reduction in Hornelen is (5°/25° * 100 =) **5 %**, and the reduction in Håsteinen (7°/35° * 100 =) **5 %**, i.e. the reduction is equal in percentage. In summary, since the “not folded” Hornelen bedding-dip is normally set to be **25°**, the “not folded” dip-value of **35°** has been used for Håsteinen in the present work.

6.4.2.3 NUMERICAL MODELLING

Introduction

In an ongoing work, the “ramp basin” model is being tested by numerical modelling. This implies modelling of the crustal-scale extensional geological development of the Håsteinen and adjacent areas. In the present thesis, the purpose of the modelling has been to see if simple models can be made that would reproduce the bedding-dips as well as the bedding/substrate angles that – influenced by isostatic movements – formed during the formation of the Håsteinen and Hornelen basins during westward movement of the Upper Plate. This modelling would also allow comparisons to be made between the two basins. However, since the purpose of the numerical models presented in the thesis is merely to show that vital geometries are in fact *possible*, the modelling of both the “extensional” and “isostatic” movements” is done with one set of values for the relevant parametres. More comprehensive numerical analysis, with testing of several values for each of the many parameters, would be beyond the scope of the thesis.

The modelling has been carried out on UNIX workstations, by means of the balancing/restoration/forward modelling program “2D-Move” version 3.00 (Midland Valley Exploration Ltd. 2000a). Interesting results have already been obtained, as illustrated in the profiles of **Fig. 6.3.a–k**, (see also **Fig. 6.1.a–n**) – and the following section will present and discuss these profiles. Although somewhat preliminary, the results show some of the potential of the ongoing modelling.

<u>*Main features of the program*</u>: The 2D-Move manual (Midland Valley Exploration Ltd. 2000b), gives the following information:

> "*2D-Move is a structural analysis and modelling program that allows line length and area balancing of cross-sections. Both structural restoration and foreward modelling can be carried out with 2D-Move. Sections from any tectonic regime can be modelled including those in extension, contraction, inversion or salt tectonics. (….). Features include depth conversion, decompaction (coupled with burial history and Airy isostacy), depth to basement, line-length and flexural slip unfolding, line and area balancing, and restore. Kinematic algorithms include Inclined Shear, and Fault Parallel Flow. Footwall deformation due to isostatic loading and unloading allows full 2D backstripping of regional sections*".

Construction of the cross-sections

In the present work, *forward modelling* has been used to model the development of the Håsteinen Devonian basin and adjacent ares. In the following description, the term "**user**" designates the person performing the modelling. When constructing an initial cross-section, various parametres, as line-lengths and -dips, may be chosen by the user. On the profiles, the "west" is to the left, and the "east" is to the right. In the following, details of the construction of <u>*one*</u> possible cross-section will be given.

<u>*Subaerial surface and detachment zone*</u>: On the PC screen, the initial cross-section is drawn with the following features: a regional horizintal line is drawn to define the subaerial surface (*upper light-blue line*, visible in e.g. **Fig. 6.3.b**). Then a detachment fault line is drawn, containing a lower flat (in the west), a ramp, an upper flat (in the east), and finally a steep line segment reaching up to the cut-off point at the "daylight" surface (*lower light-blue line*, visible in e.g. **Fig. 6.3.h**). These light blue lines will serve as reference lines that will stand in the backgroud with fixed positions on the grids of the sections. The area above the detachment is defined, for the program, as a hanging wall, and the area below as a footwall. According to the instructions given by the user, the program will make the Upper Plate move as in an extensional regime. The detachment zone in the model will then represent the Nordfjord–Sogn Detachment Zone, separating the Upper Plate from the Lower Plate. Applied to the study area, the Håsteinen basin will be formed above the *ramp* on the western (left) part of the section, whilst the Hornelen type of basin will appear above the *"listric" fault* to the east (right). To help the reader to locate features in the profiles, it has been convenient to divide each profile into four parts, which from east to west are (i) the *eastern fault*, (ii) the *upper flat*, (iii) the *ramp*, and (iv) the *lower flat*. The grid squares on the cross-section represent **10 x 10 km**, defining a stationary reference pattern that allows easy recognition of the movements.

<u>*Dip-angle of the detachment*</u>: The eastern detachment segment, i.e. the eastern fault going from the cut-off point and down to the upper flat, has been drawn with a dip of **25°** towards west. The upper flat is drawn with **0°** dip (horizontal), the ramp dips **40°** west, and the lower flat **5°** west. The geological reasons for choosing these dip values will appear below.

Thickness of the Upper Plate: As seen on **Fig. 6.3**, the Upper Plate is **10 km** thick east of the ramp. Since also the ramp itself have a vertical height of **10 km**, the Upper Plate thickness is **20 km** west of the ramp, increasing gradually westwards. If we assume a normal thermal gradient of **30°C/km**, the **10 km** depth to the the detachment zone defining the upper flat would give a temperature of **300°C/km**, roughly corresponding to lower greenschist facies. West of the ramp, the detachment zone lies at a depth of **20 km**, which would correspond to a temperature of **600°C** (when assuming the same thermal gradient), corresponding to lower/middle amphibolite facies.

Thickness of the Devonian basins: Both of the Devonian basins have been drawn with a vertical thickness of **10 km**. To the east of the ramp, the basin thickness is controlled by the thickness of the Upper Plate, and west of the ramp, the basin thickness is controlled by the height of the ramp. With a thermal gradient of **30°C/km**, the **10 km** basin thickness implies temperatures of around **300°C** at the bottom of both basins, corresponding to lowest greenschist facies metamorphism or to conditions just below this. This is compatible with the degree of metamorphism reported from the Devonian basins of western Norway (Seranne & Seguret 1987; Svendsen et al. 2001).

Westward movement of the Upper Plate

Events in the cross-sections: The start-up of the westward movement of the Upper Plate will illustrate how the program works: as a result of such movements, a *half-graben* is generated in the easternmost area, and a *ramp depression* is formed above the ramp (**Fig. 6.3.b**). For reasons explained earlier (Sect. 6.4.2.1), the ramp area will become an analogue to the Håsteinen basin, and the eastern halfgraben an analogue to the Hornelen basin. For simplistic reasons, the easternmost detachment segment that dips **25 °W**, is drawn as a *straight* line from the cut-off point to the upper flat, and not as a line with a listric shape. Consequently, a *curved* rollover anticline is not formed in the hanging wall. Generally in **Fig. 6.3**, the hanging wall *shear vector* that controles the accommodation of the hanging wall onto the subjacent detachment zone, is set to be vertical. As a geometrical necessity, the *bedding-dip angle* that develops will therefore directly reflect the dip of the eastern fault. If the fault had been drawn with a listric surface, the curvature of beds that form the roll over anticline would have mirrored the curvature of the listric fault (although the roll over curvature would have been turned upside down compared to the listric curvature).

The first four cross-sections of those displaying gelogical processes (**Fig. 6.3.b-e**), can be used to illustrate the step-by-step development in the modelling:

- In the first cross-section (**Fig. 6.3.b**), the first *extension* (“extension 1”) has been executed, resulting in westward movement of the Upper Plate, and formation of surface depressions.

- In the second section (**Fig. 6.3c)**, the first *isostatic uplift* (“uplift 1”) has been completed, an effect of the thinning of the crust caused by the preceding extension.

- In the third section (**Fig. 6.3.d**), the first *deposition* (“deposition 1”) has been carried out, in the form of a new bed with a red colour, deposited in both basins.

- In the fourth section (**Fig. 6.3.e**), this has led to the first *subsidence* (“subsidence 1”) of the crust below the basins, due to isostatic adjustments caused by the increased bed-load.

These four steps, termed “stage1”, “stage 2”, “stage 3” and “stage 4” respectively (**Fig. 6.3.b–e**), make up “cycle 1” of the modelling (**Fig. 6.3.b–e**). In the next cycle, termed “cycle 2”, the same four stages in the modelling were carried out once more, but now with deposition of a green layer. However, in contrast to “cycle 1”, where all four stages are shown in **Fig. 6.3**, only the “stage 3” section (“deposition”) is shown from “cycle 2”. The same situation also applies to the rest of the cycles (“cycle 3” to “cycle 7”): all four stages were produced in the modelling, but only the “stage 3”-sections are shown as representatives of the cycles (**Fig. 6.3.f–k**). This reduces the total number of sections shown in **Fig. 6.3,** from **31** to **11**. The selection of ”stage 3” sections is justified by the fact that these sections also incorporate and display the result of the foregoing ”extension” (“stage 1”) as well as the foregoing “uplift” (“stage 2”). Hence, the only element missing in **Fig. 6.3** is the *subsidence* occurring in the “stage 4” sections.

During the extensional movements, the Upper Plate is moved westwards, and in both basins, this causes the newly deposited bed-surfaces to move downwards. During the movement, the *Håsteinen beds* preserve their primary *depositional unconformable contact* against the west-dipping *Upper Plate* surface (of Høydalsfjorden Complex rocks). This is very different from the situastion in the eastern Hornelen type basin, where the sediments rests with a *fault (shear) contact* against the *fault*. The angular bend that has developed in the Håsteinen beds is a geometrical necessity, that forms as the beds move downwards. The bend reflects the straight ramp and the angle at the ramp-lower flat transition. A curved ramp would have given beds with a curved shape. In the cross-sections of **Fig. 6.3**, the isostatic adjustments caused by the unloading or loading of the crust, is well displayed. The uplift is particularly easy to see, as the detachment zone is lifted *above* the light-blue reference line.

Algorithms used during the modelling:

Two algorithms have been used during the modelling of the Håsteinen area: the “Flexural isostacy algorithm”, and the “Inclined shear algorithm”.

Flexural isostacy algorithm:

According to the “***2D-Move***” manual, the algorithm that calculates and performs the isostatic adjustments (*Flexural isostacy tool*) is based on the calculations of Turcotte & Schubert (1982). The tool allows several parametres to be chosen by the user, and in the present case the parametres have been set as follows (underlined):

- Load surface:	(Line in the cross-section, picked by the user).
- Load base; automatic or user-defined:	Automatic (= line below load surface)
- Load density:	**2.650 g/cm^3** (as for sandstones)

- Effective elastic thickness (EET): **1 km**.
- Mantle density: **3.300 g/cm^3**.
- Subaerial or submarine deposit: Subaerial deposit.
- Copy padding or zero padding: Copy padding (= load surface is extended to twice its length to ensure accurate results).
- Load or unload: (Choose as needed in each case).

Two aspects related to this algorithm is of particular interest; firstly the parametre “Effective elastic thickness” (EET), which is an input value for the algorithm, and secondly, the footwall uplift that appears in the models as a consequens of the calculations done by the algorithm, for this extensional setting. These two parametres are discussed in the following:

Effective elastic thickness (EET). This parametre assumes the lithosphere is isotropic and conforming to the constant values of Young’s modulus and Poisson’s ratio (Midland Valley Exploration 2000b). In practice, the parametre desides the stiffness of the lithospere when it is subjected to loading or unloading. A small value for the EET implies that a small area outside the load is effected by the vertical movements, i.e. that the vertical movements die out soon laterally, away from the load. It is difficult to set this value independently, and in the present case the simple value of **1 km** has been used.

Footwall uplift. At the *cut-off point* of the detachment zone, a footwall uplift has been formed as a result of the isostatic adjustments. Such uplifts are common at margins of extensional basins. At the *ramp area*, no marked footwall uplift appears to be produced at the advanced stages of modelling. This is due to the steeper dip of the ramp, which causes a thicker layer of Devonian sediments to be deposited after each extensional event. This causes the subsidence, resulting from deposition, to be of about the same magnitude as the uplift caused by the extension, and this prevents the footwall uplift to form.

West of the ramp, however, the crust is subjected to uplift, as seen by the detachment zone that is lifted above the light blue reference line. This uplift is a result of the westward movement of the Upper Plate, on a detachment zone that dips **5 °W**. The dip of the detachment gives the Upper Plate a “wedge shape”, and with westward movement, the thickness of the Upper Plate becomes constantly reduced, above any point in the Lower Plate. This produces an unload that causes overall crustal thinning, leading to the isostatic uplift.

Inclined shear algorithm:

This algorithm calculates and executes the westward movements of the Upper Plate during the extension, and also designs the correct geometrical relationship between fault geometry and hanging wall deformation/accomodation (Midland Valley Exploration Ltd. 2000b). Several parametres must be set by the user, and the following values have been used:

- Footwall: (Line, picked by the user)
- Hanging wall: (Line, picked by the user)
- Heave or Join beds: Heave (= allows horisontal heave to be specified).
- Horisontal heave: **-5 km** (= the amount of extension in each movement).
- Fault: (The detachment line, picked by the user).
- Objects to be moved: (Bedding lines to be moved, picked by the user.
- Autoextend fault; towards left or right: Left (= extends fault, so that Upper Plate is not moved outside the drawn fault.
- Shear angle: **90°** (= angle of shear vectors in the hanging wall. **90°** = vertical. Negative value means dip towards the right, positive values towards the left).

Two of the parametres in this algorithm are of particular interest; the *horizontal heave*, and the *shear angle* of the hanging wall. These are discussed in the following:

Horisontal heave. The value of **-5 k**m is set for practicle purposes. (The "minus" in "**-5**" indicates movement towards the left, according to program convention). If the value is selected too small, the lines representing bedding will become positioned so close to each other that they "grow together", making the cross-sections difficult to read. The selected **-5 km** is a compromise value that gives good readability. In geological terms it is of course unrealistic with **5 km** extension in one event, but in the overall picture, the value does not make the model unrealistic. The alternative, which would be to perform the extension "a little by little" (low heave value) would lead to deposition "a little by little", but would not severely change the final amount of sediment deposited in the basins. Therefore, large extensional heaves, such as **5 km,** does not invalidate the model, given the present purpose of the modelling.

Shear angle. This is an important parametre in the Inclined shear tool, as it defines the orientation of the shear vectors controlling the "collapse" of the hanging wall elements down onto the fault plane, removing the "void space" that would otherwise appear. A shear angle of **90°** gives the simplest situastion, and modelling commonly starts with this value if a more exact value is not apparent (Midland Valley Exploration Ltd. 2000b). In the present modelling (**Fig. 6.3)**, the value has been set to **90°**.

Compaction/decompaction algorithm: not applied

The 2D-Move program has a compaction/decompaction algorithm, and corresponding tool. The Håsteinen Devonian sediments experienced some degree of compaction during the burial, but the amount of compaction is assumed to have been very limited. This is based on the fact that the Devonian sediments in Håsteinen consist of very poorly sorted debris flow deposits, containing all grain fractions up to boulder. The clasts/grains have therefore been very closely packed during deposition, leaving a limited pore volume. Quartz cement has filled the pores, and later pressure solution has been limited. Hence, everything considered, it has been found unnecessary to calculate the compaction.

Bedding and detachment geometries, as controlled by the 90° hangingwall shear-angle and the isostatic movements

As mentioned above, an interesting geometrical relationship exists between the *detachment geometry* and the *bedding geometry* **–** on the conditions that the shear angle in the hanging wall is **90°,** and on the condition that the bedding geometries have not been severely altered by e.g. isostatic movements**:**

> the shape and dip of the Devonian beds will perfectly mirror the shape and dip of the detachment zone that exists below the sediment-receiving basin (albeit with the bedding curvature being turned upside down compared to the detachment curvature).

This geometrical relationship is illutrated in the idealised cross sections of **Fig. 6.1**, where the bedding geometry in the Hornelen type basin is controlled by the *easternmost fault segment*, and where the bedding geometry of the Håsteinen basin is controlled by the *ramp* (/Upper Plate). In **Fig. 6.1**, the isostacy algoritm has not been applied in the modelling, as focus is merely on the basic geometries of the detachment system. In **Fig. 6.3**, however, the modelling includes isostatic movements, and we see that the simple geometrical relationship is actually somewhat altered by the isostatic *uplift*, although the alteration of dip *angles* is *small* in the Håsteinen case and in fact *absent* in the Hornelen case.

At the Hornelen type of basin in the east (**Fig. 6.3**), the isostatic uplift does not lead to a change of the geometry (dip/shape) of the subjacent detachment fault, and the geometrical relationship between detachment and bedding is maintained. Since the bedding in the present-day Hornelen area dips **25°** towards east, the modelling must be able to reproduce this situation. By selecting exactly the **25°** *westward* dip for the eastern fault, the Devonian beds will in fact **–** due to the **90°** shear vector in the hanging wall **–** be rotated to the correct **25°** *eastward* dip during the extension.

For the Håsteinen basin, it is assumed that the bedding had an eastward dip of around **35°** prior to folding, i.e. that the extensional movements down the ramp produced the **35°** eastward dip of the bedding (see Sect. 6.4.2.2). In the numerical modelling, it must therefore be possible to reproduce this bedding-dip of **35°**. In **Fig. 6.3.a**, the ramp is drawn with a dip of **40°W**, and the lower flat is drawn with a dip of **5°W**. With a **90°** shear vector in the hanging wall, the combined **40°/5°** dip of the ramp/lower flat will, as a geometrical necessity, rotate the Devonian beds precisely to the desired **35°** eastward dip (although the angle becomes slightly altered due to the isostatic movements). After movement of the Håsteinen beds and Upper Plate down the ramp, the beds will eventually stand unconformably with approximately the **35°** dip-angle against the subhorisontal Upper Plate. Just west of the ramp, isostatic adjustments have led to uplift, which has made the detachment zone subhorisontal, and this has increased the bedding-dip in Håsteinen with up to **5°**. West of the Håsteinen basin, the detachment zone has preserved its dip of **5°**.

Westward movement of the Upper Plate, as influenced by the isostatic uplift

The ability of the Upper Plate to move westwards on the detachment zone, relative to the Lower Plate, is controlled by the presence of a minimum stress (σ_3) oriented horisontally in a roughly E-Wdirection, and by the degree of friction along the detachment zone. The E-W oriented least stress would be consistent with the divergent motion of the Baltic and Laurentian cratons that started with the **Mode-I** movement of the entire Caledonian orogenic wedge at **~402 Ma** (Fossen & Dunlap 1998). Thus, a stress situation generating extension seems well documented. The matter of friction, however, should be briefly considered: With a minimum stress (σ_3) oriented as described, the movement will take place as long as the detachment zone is oriented with a slight westward dip and as long as the friction is overcome, for example due to high fluid pressures in the detachment zone. In **Fig. 6.3**, the lower flat (west of the ramp) is initially drawn with a dip of **5 °W**, while the upper flat (east of the ramp) is drawn horizontal, **0°**. For the model to be geologically "correct", the upper flat should have had a westward dip of a few degrees instead of being horisontal. However, the combination of **5°** dip of the lower flat and **0°** of the upper flat in Fig. 6.3 has been chosen to make it possible to test differences in the development of the two settings. At the Hornelen type basin, the isostatic uplift in itself produces an up to **6°** westward dip of the upper flat of the detachment zone. This feature may indicate that, once the basin has formed, the westward movement of the Upper Plate and basin sediments may be facilitated by the increased dip of the detachment zone. In **Fig. 6.3.k**, the maximum vertical uplift for the "Hornelen" area is **~2.5 km**.

At the Håsteinen ramp basin, however, the situation appear to be the opposite, as the initial dip of **5°W** for the detachment zone (lower flat) is reduced to **0°**, during the movement of the Upper Plate. In **Fig. 6.3.k**, the uplift was measured to **~2.5 km**, but would apparently have increased with further extension. As mentioned above, the uplift of the entire crustal length west of the ramp is caused by the wedge-shape of the Upper Plate, a shape that reduces the crustal thickness during the movement. West of the ramp in **Fig. 6.3**, within the area with **0°** dip of the detachment, the Upper Plate would suffer the risk of being "torn apart" during the westward movement. It appears, therefore, that the Upper Plate, to be able to continue as one unit, would benefit from a "push" from its thin part east of the ramp. If the upper flat, east of the ramp had been drawn with a dip of a few degrees, instead of being drawn with the horisontal orientation, the Upper Plate would also here have had a "wedge-shape" that would have given uplift of the same type as at the lower flat. This uplift would have been combined with the uplift already controlled by the formation of the "Hornelen-type" basin, and would possibly have secured a continuous westward dip of the upper flat detachment zone. The general uplift of the *upper flat* due to the wedge-shape of the thin Upper Plate here, would also have given uplift of the ramp area, since the Upper Plate passing above it would gradually become thinner.

Total isostatic uplift: an effect of Western Gneiss Region exhumation, as well as detachment movements and basin formation

It is worth noting that below the Håsteinen-type basin, where the detachment zone has been uplifted and rotated to a near horizontal position, and further to the west where the entire Upper Plate is uplifted – the same uplift has also affected the Lower Plate and the Nordfjord–Sogn Detachment Zone, i.e. the entire crust. Also the uplift east of the ramp, including the “Hornelen-type” basin, will give a general crustal uplift. This is interesting, because it could mean that the uplift – and thus the corresponding metamorphic retrogression recorded in the NSDZ during the coeval extension – occurred partly as an effect of the isostatic movements imposed by detachment-dip and basin formation, and not purely as a result of the general exhumation of the eclogitic Western Gneiss Region. Forthcoming models will show if this development prevails as the different parametres are changed.

6.4.2.4 "LOWER PLATE INCISEMENT" MODEL

Outline of the Lower Plate incisement model

The data recorded in the HDM can also, theoretically, be explained by a different model which has been termed the “lower plate incisement” model. In this model, rock "flakes" that have been incised from the Lower Plate in the east, are transported down along the listric fault that defines the easternmost part of the detachment zone. The model is illustrated in **Fig. 6.5.a-e**, and can be explained as follows: the crustal-scale extension has lead to generation of a west-dipping listric fault that at depth continues as a subhorizontal detachment. The first event of west- and down-movement of the Upper Plate has created a depression formed as a half graben (**Fig. 6.5.a)**. In the depression, sediments are then deposited with a *depositional unconformable* contact against the west-dipping *fault zone* (detachment zone), which defines the eastern basin margin. If the fault zone is steep, the angle will be large between the bedding and the fault zone, and the fault zone itself will constitute the Devonian/substrate contact surface. To preserve the primary unconformity here, movements must not occur along the fault during or after deposition, as this would destroy the depositional unconformity, particularely when cementation has “welded” the sediments to the fault. Hence, a phase of active tectonics, with rapid down-movement combined with limited sedimentation, must subsequently be replaced by a tectonically quiet period with active sedimentation and no fault movements. Later, when a new phase of extensional movement is again about to take place, a completely new fault plane have to be developed *east of the pevious one*, within the Lower Plate (**Fig. 6.5.a)**, so that the primary unconformity established at the foregoing fault, in the quiet period, is not destroyed. With the second fault present, the first proper "flake" of Lower Plate rocks with the primary unconformity preserved on the western side, will then be incised and carried down the fault plane (**Fig. 6.5.b)**. When the area again becomes tectonically quiet, sediments are once more deposited with a primary unconformity towards the newly formed fault plane. This sedimentary sequence is related to the movement of the incised “flake”, and may therefore be called a sedimentary “incisement sequence”. It is worth noting that an additional depositional unconformity is, from now on, present also on *top* of the incised “flake”.

Subsequently, the same development, with alternating incisement faulting and deposition is repeated over and over again, and both the bedding-surface and the Devonian-substrate contact surface will constantly rotate to form the roll-over anticline, as the bedding is moved down the listric fault. When the sub-horizontal base of the listric part of the detachment is reached, the east-dipping Devonian bedding will eventually be standing with a high angle towards substrate rocks that have a sub-horizontal topographic envelope surface (**Fig. 6.5.e)**. At an even more advanced stage, the same bedding/substrate geometry can be seen to continue westwards (**Fig. 6.5.e)**. The repeated generation of new “flakes” will secure that the Devonian sediments are always deposited on the rocks of the Høydalsfjord Complex, which is consistent with the situation in the field. (If movements could continue unlimited on one single fault, the Håsteinen sediments could eventually be resting with a shear contact against the subjacent mylonitic gneisses of the Eikefjord Group i.e. the Nordfjord–Sogn Detachment Zone, which would be inconsistent with field observations).

In the litterature, transport of wedges of “fault blocks” down along detachment faults have been described from the Basin and Range, USA, by e.g. Wernicke (1981), terming them "rock sheets" or "nappes"; and by Lister & Davis (1989), terming them "incisement nappes". Also Gibbs (1984) described wedge-shaped bodies, and named them extensional “riders”. The series of wedges were shown to have moved down along the datachment faults to form a domino-style succession of wedges resting on the detachment zones. Hence, these studies indicate that series of “flakes”, wedges, or fault blocks can indeed, once formed, move down along detachment faults.

Discussion of "Lower Plate incisement" model

From the descriptions above, it appears that the incisement model may ***theoretically*** explain the data in the HDM. In the following, it will therefore be closer considered whether the model may work in practice, and arguments in favour of and against the model will be presented.

In favour of the model. The model implies that the Devonian basin is formed in a depression (half-graben) located on top of the easternmost part of the Upper Plate. At the eastern boundary of the half-graben, the sediments rest against the fault surface of the Lower Plate. In these respects, the model is compatible to the "Hornelen type" setting, and this may be an argument in favour of the model. However, the model differs from the "Hornelen type" setting by one important feature: in the incisement model, the deposition of sediments occur unconformably against *inactive* "old" fault planes (which are preserved by down-movement of "flakes" of the Lower Plate), instead of deposition directly towards the *active* detachment zone as in the Hornelen Devonian Massif. The incisement model does explain the rapid, coarse sedimentation, since the basin would benefit from the generation of relief to the east due to the uplift of the Lower Plate.

Against the model. The model leads to several problems that make it seem less likely as an alternative explanation of the HDM-formation, and some of these problems are briefly reviewed below:

(1) In the Håsteinen area, it is very likely that the constant bedding-orientation that has been documented within each of the two limbs of the Osstrupen syncline, continued further westwards. Thus, for the model to work, the entire Upper Plate in the Håsteinen area and the areas further west would have to consist exclusively of down-moved "flakes" of Lower Plate rocks, separated by discrete shear zones. Although the presence of such "flakes" are difficult to test due to the lack of structural markers in the Høydalsfjorden Complex (i.e. lack of laterally persistent lithological layers), it is considered very unlikely that the area containing Høydalsfjorden rocks, presently measuring at least **20 km** in the W-E direction, should consist exclusively of such "flakes".

(2) It is difficcult to envisage how a repeated geneation of *new* listric incisement faults in the Lower Plate could have been accomplished once the Nordfjord–Sogn Detachment Zone, with its eastern listric fault segment, had been formed. The westward movement of the *already* formed Upper Plate of the model would probably accommodate the extensional stress of the system, and it would then be mechanically unfavourable to constantly form new faults eastwards in the Lower Plate.

(3) It is very likely that the "old" fault planes would be *reactivated*, leading to destruction of the depositional unconformities.

(4) The "flakes" would have to be deeply rooted – in the sense that they would have to continue down to the level where the listric part of the detachment zone flattens out. In addition, also the *thickness* of the “flakes” would have to be largely maintained down to the flatter part of the detachment. If these conditions are not fulfilled, the dip of the eastern listric fault plane would soon become more and more reduced, and the plane would eventually approach a sub-horizontal orientation. This would make incisement of new “flakes” even less likely, and moreover reduce the thickness of the Devonian basin. The condition, that thickness and depth of the “flakes” must be largely maintained are problematic.

(5) In the model, the substrate rock surfaces that form the unconformable contacts below the Håsteinen deposits would essentially be abandoned fault planes. In the studied field, however, the primary unconformities display an irregular eroded palaeosurface (local topopraphy) which does not appear to be compatible with this. The model also offers the latter type of unconformity, present on the "top flats" of the subdued "flakes", where sediments would be resting on an irregular eroded palaeosurface and not against a palaeofault. In theory it is possible that all the primary contacts exposed in the HDM are from such "top flats", although this is considered higly unprobable since the palaeofault contacts are, by proportion, the dominating type in the model.

(6) It is unlikely that the fault movements could accumulate to very large throws without simultaneous sedimentation. To be able to preserve the unconformities that are developed on the abandoned fault planes, as a result of continous sedimentation, new faults would constantly have to be generated to the east of the previous one. Generation of such faults would be necessary even in situations where the movements on "old" faults were small. It would here not be possible to argue that sedimentation occurred only, or mainly, on the <u>top</u> of the "flakes" instead of towards the abandoned fault planes, since this would "remove" the faults and thereby the basin-forming process. One single incisement event, e.i. the downward movement of one particular

“flake” would in practice occur by incremental accumulation of several smaller fault movements. If the downward movement of *one* single “flake” lead to formation of high fault scarps, and if the fault thereafter became tectonically quiet and gave way to rapid deposition, then the whole sedimentary “flake sequence” that formed in relation to this particular “flake” would have essentially parallel internal bedding, since no growth-faulting would occur. If also the next new fault to the east developed a high fault scarp before sedimentation started, an angular relationship would be formed between the previous sedimentary sequence – which would now have a gently easward bedding-dip – and the new horizontally deposited sequence (e.g. **Fig. 6.5.c**).

As illustrated by the above discussion, several problems are attached to the model. Furthermore, the field work in the rocks of the Høydalsfjorden Complex, which is situated below the Håsteinen Devonian Massif (HDM), has not provided evidence in support of the model. One could argue that it would be a difficult task, in the field, to reveal whether the Høydalsfjorden substrate (Upper Plate) is made of such "flakes" of down-faulted lower-plate rocks, or not. This difficulty could be said to arise from the presence of top-to-the-west extensional $\mathbf{D_3}$-mylonites that preceded the formation of the Håsteinen basin, the argument beeing that the $\mathbf{D_3}$-shear fabric locally could partly be a result of “flake”-shear that has been mis-interpreted as $\mathbf{D_3}$-shear. However, the $\mathbf{D_3}$-mylonites are located just below the Håsteinen deposits. It is implisit in the incisement model that “flake”-fault zones located so close to the Devonian deposit would reflect very shallow crustal levels at the time of their formation, implying that the “flake”-zones would be relatively thin and that any fault “rocks” would be fault gouge or breccia. It is also likely that potential “flake”-zones would have to cut trough the $\mathbf{D_3}$-mylonite foliation with some angle. Furthermore, the movement of the “flakes” down the listric fault would lead to rotation of the $\mathbf{D_3}$-mylonite foliation within the “flakes”, for example from a horizontal orientation to an eastward dip. $\mathbf{D_3}$-mylonite zones would therefore tend to show a mismatch across the “flake” fault zones, due to the displacement. In the field, however, the $\mathbf{D_3}$-mylonites have a very characteristic appearance and lateral continuity, and no observations have been done that could indicate presence of “flake”-fault zones. Also, along the northern margin of the HDM, where the Devonian/substrate contact may be traced precisely – regardless of whether it is directly exposed or not – the contact runs with a stable trend and without noticable step-like offsets that could indicate “flakes”. (The few recorded faults with minor offsets are later features, and not relevant to the present discussion). Eventually, it should be reminded that the $\mathbf{D_3}$-mylonites are cut by the unconformity below the HDM, showing that the mylonites formed prior to basin formation. Although the incisement model have some attractive elements, the problematic sides of the model are by far the more prominent, raising questions as to whether the model could work in practise. Accordingly, it is here suggested that the model be considered unlikely.

6.4.2.5 "PALAEO-MOUNTAIN" MODEL

The sedimentary beds in the HDM have been deposited sub-horisontally, and with a primary unconformable contact against the substrate rocks. As discussed in Sect. 5.5.8.6, bedding along the axial trace of the Osstrupen syncline dips **53°ESE** and impinges against a surface of substrate rocks that is sub-horizontal. This results in an angular relationship between bedding and the substrate rocks along the axial trace profile

(**Plate 2**, **Fig. 5.86.a**). Rigid back-rotation of bedding from the **53°-**dip to a horizontal orientation, which implies a similar rotation of the attached substrate, would cause the unconformable Devonian/substrate contact to form a moderately to steeply dipping palaeo-slope (**Fig. 5.85**). The stratigraphical thickness measured along this slope would be minimum **5.8 km (Fig. 5.85)** – probably **11.2 km** – and possibly even larger. At the time of deposition, the heights of mountains and topographic reliefs in these parts of the Caledonian mountain chain were probably near the highest values possible for such mountain chains. It is diffcult to envisage that Devonian sedimentary beds were first deposited against the slopes of such high mountains, and later rotated to achieve the present **53°ESE** dip of the bedding. In particular, the stratigraphical thickness in the HDM, along the axial trace, is considered too large, and the rotation needed, too unrealistic, to make such a model likely.

Hence, since an *en block* rotation of the Håsteinen beds must be considered unrealistic **(Fig. 5.85)**, the model must be rejected.

6.4.2.6 CONCLUDING REMARKS

Both the "ramp basin" model and the "Lower Plate incisement" model succeedes in explaining the major data that are related to the formation of the HDM. However, as shown above, a number of problems are related to the the "Lower Plate incisement" model. Everything considered, the “ramp basin” model therefore offers the best explanation for the development of the Håsteinen area.

6.4.3 ORIGINAL SIZE OF THE HÅSTEINEN DEVONIAN BASIN

South of the Håsteinen Devonian Massif (HDM), the north-dipping mylonite belt of the Standalen segment of the Nordfjord–Sogn Detachment Zone (NSDZ), separates the HDM and the Høydalsfjorden Complex to the north from the eclogitic gneisses (Western Gneiss Region, WGR) to the south. In accordance with the detachment model, the Standalen segment of the NSDZ could be the original southern margin of the HDM. West and north of Håsteinen, however, the positions of the original margins are more uncertain. It is likely that the basin originally continued further towards the west, but it is uncertain whether the sediments in these more western areas, did coalesce with the once much larger Hornelen and Kvamshesten basins.

Towards the north, it is possible that the Håsteinen basin ended near the present Sunnar Fault, which now separates the HDM and the Høydalsfjorden Complex to the south, from the Eikefjord Group mylonites (NSDZ) to the north. Alternatively, the basin might have ended further north against a fault developed within the now eroded rocks that were once situated *above* the mylonitic detachment zone present between the HDM and the Hornelen DM. Thus, the position of the original northern margin of the HDM is uncertain at the present stage.

The issue of whether the Håsteinen and Hornelen basins were connected –possibly at higher stratigraphic levels – cannot be answered with certainty. However, some factors of relevance can be mentioned: The erosion that has reduced the thickness of the NSDZ between Håsteinen and Hornelen has apparently removed only a modest part of the zone, indicating that the original detachment zone reached only slightly higher than the presently exposed level (for example maximum **1 km** higher). This implies that during the extensional movements, the Upper Plate was positioned just above the presently exposed mylonites. Sediment transport directions in the Hornelen Devonian Massif show that the Upper Plate source area was clearly present to the south of Hornelen, i.e. between the Håsteinen and Hornelen basins. This is documented by the presence of offset-stacked marginal alluvial fans along the southern margin of the Hornelen DM (e.g. Steel et al. 1985), built out northwards into the basin, from the southern source area. Each deposited layer was rotated to an eastward dip after rollover on the listric fault, as the basin moved westwards, giving space for new layers supplied from the south. This implies that Hornelen was a separate basin at the time when these fans were deposited at the basin surface. Therefore it is not likely that the Håsteinen and Hornelen basins coalesced, at just slightly higher levels in the original vertical stratigraphy. (See also discussion at Sect. 6.4.4).

Since the palaeo-transport directions for the HDM are not known, however, it is not possible to use the transport direction of aluvial fan conglomerates along the northern HDM margin to decide whether this margin is the original basin margin. And, since the Lower-Plate-source-rocks for the HDM-sediments have been removed from the source areas due to the tectonic and erosional off-stripping of the Lower Plate-areas of the system, transport directions and basin margins cannot be inferred from such source areas either. It must thus be concluded that (1) the Standalen segment of the NSDZ represents the original *southern margin* of the HDM, (2) the basin probably continued for large distances westward on the Upper Plate, and (3) the Upper Plate rocks that were situated on top of the detachment zone in the north, probably separated the Håsteinen basin from the Hornelen basin, at least during deposition of the Hornelen south margin fans, although it cannot be excluded that the Håsteinen and Hornelen basins coalesced towards the west. Future investigations on palaeo-current directions in the northern part of the HDM should reveal whether sediments were supplied from the north, i.e. whether the now removed Upper Plate, that were present on the detachment zone in the north, represented an original basin margin that supplied the alluvial fans of the Vikafjell formation.

The sedimentation of the Devonian basins is a result of the combination of uplift of the *source area* represented by the Lower Plate, and relative lowering of the depositional basin areas on the Upper Plate. It should be noted that at the time of deposition of the Håsteinen sediments, the source rocks of the Lower Plate was of the *same type* as the rocks now constituting the Upper Plate, i.e. of the same type as the rocks constituting the substrate to the Devonian sediments.

6.4.4 DEVELOPMENT OF THE DETACHMENT ZONE (EIKEFJORD GROUP)

In the present section, five subjects are discussed: **1)** The tectonostratigraphic position of the rocks of the Eikefjord Group, i.e. the Nordfjord–Sogn Detachment Zone (NSDZ), in the study area. **2)** The geological history that preceded the formation of the S_2-mylonite of the Eikefjord Group. **3)** An apparently *paradoxical* situation arising from the location of the Nordfjord–Sogn Detachment Zone between Hornelen and Håsteinen, where the paradox is as follows: could a *Lower Plate source rock* area have a *fixated* position south of Hornelen and deliver sediments to a comparably westward-moving Hornelen basin/Upper Plate – when the presence of detachment mylonites between Hornelen and Håsteinen suggests that the ***Lower Plate source area*** did ***itself*** move westwards on the mylonites – implying that this Lower Plate was rather an *Upper Plate*? **4)** The apparent absence of NSDZ-type mylonites in the Høydalsfjorden Complex. **5)** The depth from the Upper Plate to the NSDZ in the study area, as revealed by brittle faults along the Florø peninsula horst.

Tectonostratigraphic position of the detachment zone

Generally, the tectonostratigraphic position of the rocks effected by the Nordfjord–Sogn Detachment Zone may reveal whether the detachment zone has cut the regional tectonostratigraphy or not. The tectonostratigraphic position of the Eikefjord Group has been interpreted to be analogous to the position of the Dalsfjord Suite of the Dalsfjord Nappe (Bryhni & Sturt 1985; Andersen & Jamtveit 1990; Furnes et al. 1990; Osmundsen & Andersen 2001). In the area between Førdefjorden and Dalsfjorden, i.e. the area hosting the Kvamshesten Devonian Massif, the detachment zone is developed *below* the Dalsfjord Suite, which itself has not been affected by the detachment mylonitisation, except along its base. In the Eikefjord area, however, the igneous rocks of the Eikefjord Group are penetratively mylonitised, thus constituting the detachment zone. If the tectonostratigraphic correlation of the Dalsfjord Suite and the Eikefjord Group is correct, this shows that the detachment zone is cutting *up-section* from S to N, i.e. that the detachment zone is situated tectonostratigraphically higher in the Håsteinen area than in the Kvamshesten area. This indicates that the detachment zone cuts through the tectonostratigraphy, and that it does not merely reactivate old contractional thrust zones. The omission of structural section supports the model suggested by Fossen (1992, 2000), who established the term **Mode-I** for the westward movement of the orogenic wedge on the decollement zone, and the term **Mode-II** for the subsequent movement on the NSDZ. Also the works by Milnes et al. (1988; 1997), based on structural logging of the north Sognefjorden shoreline, supports the interpretation that the NSDZ is cutting up-section. The latter authors consider the Nordfjord–Sogn Detachment Zone as a second-stage extensional feature that cut the nappe pile after a first stage of back movement (extension) of the entire Caledonian orogenic wedge on the thrust décollement.

An alternative interpretation may be appropriate when the metamorphic development is considered. The Dalsfjord Nappe has occupied a shallow-crustal position (lower greenschist facies) during the Caledonian cycle (Andersen et al. 1998), whereas the Eikefjord rocks have experienced an amphibolite facies metamorphism. This shows that the Eikefjord/Lykkjebø rocks may have been brought to substantially deeper levels than the Dalsfjord Nappe during the Caledonian contraction, opening for the possibility that the two areas did not have the same tectonostratigraphic positions. The results of Johnston et al. (2007b), that the Eilefjord and lykkjebø rocks experienced upper amphibolite facies at **13–18 kbar**, **537–618°C**, **45–60 km**. in the time interval **425–410 Ma**, may be taken to support this interpretation.

Pre-S_2-mylonite history in the Eikefjord Group

In the present work, the metamorphic history of the S_2-mylonite of the Eikefjord Group has been interpreted to be retrogressive, and thus reflecting the crustal-scale exhumation occuring during the Devonian extensional movements. A large number of works report a similar retrogressive metamorphism in the rocks in both the detachment zone and the Lower Plate elsewhere in these parts of western Norway (e.g. Andersen & Jamtveit 1990; Chauvet et al. 1992; Wilks & Cuthbert 1994; Young et al. 2007; Johnston et al. 2007b). In the *study area,* the "oldest" **pre-S_2**-mylonite history in the Eikefjord Group is defined by an amphibolite facies mineralogy which has been suggested to be of late-/post-Scandian age (**Sect. 3.5**) and which represent the highest metamorphism documented in the study area. The possible origin of this amphibolite facies will be discussed in the present section. The four following alternatives exist for the origin of the amphibolite facies metamorphism: **(1)** the amphibolite facies mineralogy was produced by Caledonian "Jotun Nappe type" retrogression from an original Precambrian granulite facies mineralogy, **(2)** The amphibolite facies mineralogy was developed as part of the prograde subduction process that formed the Scandian high-grade assemblages elewhere in the WGR, implying that the Eikefjord Group rocks were partly involved in the subduction below Greenland (Laurentia), **(3)** the amphibolite facies mineralogy was produced by retrogression from Scandian high-grade (eclogite facies) rocks, i.e. related to exhumation of the unit, **(4)** the rocks of the Eikefjord Group developed the amphibolite facies mineralogy in a prograde process due to the vicinity to underlying high-temperature eclogitic rocks of the Lower Plate that became juxtaposed with the NSDZ mylonites during the Devonian large-scale extension. The four alternatives will first be explained individually, and then discussed jointly.

Alternative 1: amphibolite facies was caused by Caledonian "Jotun Nappe type" retrogression. According to Bryhni et al. (1981), the Eikefjord Group originally had a Precambrian granulite facies mineralogy. To illustrate the Precambrian metamorphism, a comparison can be made with the Jotun Nappe and the Dalsfjord Nappe which have been correlated with the rocks of the Eikefjord Group. In the Jotun Nappe and the Dalsfjord Nappe, this high-grade mineralogy has been retrograded via amphibolite facies and locally further towards greenschist facies during the Caledonian contractional orogenic cycle (Milnes et al. 1997, Corfu & Andersen 2002). These two nappes have been situated in a shallow crustal position in the Caledonian nappe stack during the whole Caledonian contractional orogenic cycle, and this led to the metamorphic retrogression. Thus, the Jotun Nappe and the Dalsfjord Nappe have not been subducted in the A-type continental subduction that has generated the Caledonian high-grade (eclogite) assemblages now present in the Lower Plate gneisses (WGR) found to the east of and between the Devonian Massifs of west Norway, and particularly north of the Hornelen massif.

Alternative 2: amphibolite facies was formed as a prograde assemblage due to subduction below Laurentia. In addition to the *Precambrian* high-grade assemblages that can be found for example in the Eikefjord Group and the Lower Plate (Western Gneiss Region), *Scandian* eclogites and related rocks with high-grade assemblages are generally abundant in the Lower Plate as well as in the *lower parts* of the **2–3 km** thick detachment zone. These eclogitic high-grade rocks have been interpreted as related to subduction of the Western Gneiss Complex below the Laurentian craton during the Scandian continent-continent collision. Although the high-grade rocks reported by Bryhni et al. (1981) in the Eikefjord area are probably Precambrian and not

Caledonian, the amphibolite facies mineralogy in the Eikefjord Group was possibly formed as a prograde metamorphic (“on the may down”) assemblage during the *Scandian* subduction event. This would imply that the Eikefjord rocks were partly subducted along with the WGR, i.e. that the Eikefjord rocks were not a Caledonian nappe occupying a shallow crustal position during the Caledonian orogeny. An Eikefjord amphibolite facies mineralogy related to contraction/subduction would be compatible with results obtained by Johnston et al. (2007b), who, on the basis of radiometric dating and P/T analyses, suggested that an (upper) amphibolite facies in the Eikefjord/Lykkjebø Groups occurred at **13–18 kbar**, **537–618°C**, **45–60 km**, during in the full time interval **~425–410 Ma** (Sect. 1.5 and 6.2.1).

Alternative 3: amphibolite facies was formed as a retrograde assemblage due to exhumation from below Laurentia. Also this alternative assumes that the Eikefjord Group rocks were involved in the subduction of the WGR below Laurentia. Then, after maximum subduction, the Devonian uplift led to extensive retrogression of the Scandian high-grade rocks. The amphibolite facies assemblage could represent a stage in this continuous retrogression process, and would then not necessarily represent the foregoing peak metamorphism. A possible example was obtained by Johnston et al. (2007b), who suggested that NSDZ shear in the Eikefjord/lykkjebø rocks (at lower amphibolite to greenschist facies) occurred at **8–12 kbar**, **519–641°C**, **30–40 km** – estimated to have occurred in the time interval **~410–400 Ma** (Sect. 1.5 and 6.2.1). (Note that this time interval is considered to be somewhat early (old) compared to the documented initiation of the NSDZ/Mode-II shear movement at, or just before, **394 Ma**, i.e. at the later stage of the Mode-I interval of **402–394 Ma**, see Fossen & Dunlap 1998).

Alternative 4: amphibolite facies was formed due to heat from underlying eclogitic Lower Plate (WGR). Since the Eikefjord Group constitutes the Nordfjord–Sogn Detachment Zone in the study area, the detachment rocks have been situated very close to the subjacent high-temperature eclogitic rocks of the Lower Plate that was rising from below during the Devonian crustal-scale extension. This means that the amphibolite facies mineralogy of the Eikefjord rocks could have been formed partly as a result of the juxtaposition with these high-temperature eclogite rocks.

Discussion of the four alternatives. If it had been possible to relate the Eikefjord Group high-grade relicts (granulite facies) found outside the study area, to the *Scandian* metamorphic event (*Alternative 2*), it would mean that the Eikefjord Group and the Lykkjebø Group had been "subducted" to considerable depths during the continent-continent collision. Discoveries of eclogites would have opened for this possibility. Two large bodies of gabbroic rocks (measuring **~10x1 km** and **~6x2 km**) that are present in the study area and further eastward in the Eikefjord valley, and several amphibolites, are rock types that would have been suitable for eclogite formation. However, no remnants of eclogitic rocks have as yet been found between the Håsteinen and the Hornelen Devonian Massifs, and neither have eclogites been found further east in the large area occupied by the sheared detachment zone rocks. This may suggest that the area has not been fully "subducted". However, the amphibolite facies mineralogy in the Eikefjord/Lykkjebø rocks may have formed during subduction of the *subjacent* eclogitic WGR rocks (*alternative 2*). As mentioned above, this is compatible with the report of Johnston et al. (2007b), advocating that an upper amphibolite facies metamorphism occurred at **13–18 kbar**,

537–618°C, **45–60 km**, in the time interval **~425–410 Ma** (Sect. 1.5 and 6.2.1). In thin section studies, unaltered ***prograde*** amphibolite facies fabrics have generally ***not*** been observed, as all fabrics are affected by the retrogressive processes. Nevertheless, a prograde amphibolite facies (*Alternative 2*) may have been formed during Scandian contraction and subdjacent subduction of the WGR.

As documented in the present thesis (Ch. 3.5), the rocks of the Eikefjord and Lykkjebø Groups display evidence of strong retrogradation from amphibolite facies to greenschist facies assemblages (*Alternative 3*). These retrogressive processes reflect exhumation of the NSDZ from deep to shallow crustal levels, during the top-to-the-west shear movements. As mentioned above, Johnston et al. (2007b), reported retrograde shear at **8–12 kbar**, **519–641°C**, **30–40 km**, i.e. at lower amphibolite to greenschist facies – that was estimated to have occurred in the time interval **~410–400 Ma** (Sect. 1.5 and 6.2.1). Although the age interval of the shear movements is somewhat old compared to the timing of the NSDZ as a Mode-II zone, the P/T data are consistent with the observed retrograded amphibolite facies mineralogy. The observed prominent *greenschist facies* shear fabric, however, presumably occurred at crustal levels *shallower* than the **30–40 km** estimated by Johnston et al. (2007b). The observed amphibolite facies may have formed during retrogressive prosesses (*Alternative 3).*

As mentioned above, the rocks of the Eikefjord Group have been correlated with the Dalsfjord Nappe, which was located in a shallow crustal position during the maximum Scandian collision. This correlation would imply that also the Eikefjord rocks were situated in the Upper Plate position, and that the Nordfjord–Sogn Detachment Zone was developed in the Eikefjord rocks later. In this scenario, the amphibolite facies mineralogy would have to be either a remnant of a Caledonian "Jotun-type" shallow- to middle-crustal retrogressive amphibolite facies (*alternative 1*), or the mineralogy could be a result of the close spatial relationship with the Lower Plate eclogitic rocks that were rising from below during the development of the detachment zone (*alternative 4*). In the first of these interpretations (*alternative 1*), it may be a problem that the rocks of the *Upper Plate* (e.g. the Dalsfjord Nappe) between Sogn and Nordfjord, have generally experienced a prograde Scandian metamorphism of maximum middle greenschist facies, i.e not reacing the amphibolite facies present in the Eikefjord Group. The last one of these two alternatives (*alternative 4*), i.e. that the amphibolite facies was partly developed due to heating from eclogites present in the WGR rocks rising undferneath the NSDZ, is supported by the fact that eclogites of the Lower Plate are indeed present not far from the Eikefjord Group**:** to the west of the study area, such eclogites have for example been reported from gneisses on the island of Kinn (**Fig. 2.3**) (Kolderup 1928), **27 km** west of the study area (and **12 km** west of Florø town), and also from the outskirts of the Florø town itself, **15 km** west of the study area. To the south of the study area, such eclogites have been reported from the areas to the south of the Standalen segment of the NSDZ, along the southern margin of the Håsteinen Devonian Massif (Andersen & Jamtveit 1990; Kildal 1970, Krabbendam & Dewey 1998).

The re-appearance of Lower Plate eclogites to the west of the study area, at Kinn and Florø, indicates that the eclogites are present underneath the whole NSDZ and Upper Plate, in an area encompassing at least the Hornelen sediments and the Bremanger area rocks, and the Håsteinen sediments and the Høydalsfjorden Complex. This indicates that eclogite rocks are located very close to the Eikefjord Group, making it possible that the formation of the amphibolite facies mineralogy in the Eikefjord rocks was affected by the uplift of these subjacent, high-temperasture eclogitic rocks. Until recently, no data has existed that could reveal wether the amphibolite facies mineralogy represented the peak of metamorphism, or whether it was rather only a stage in a post-peak, continuous retrogression related to the uplift. Johnston et al. (2007b), however, interpreted their amphibolite facies metamorphism of **13–18 kbar**, **537–618°C**, **45–60 km**, in the time interval **~425–410 Ma**, as represnting the *peak* of metamorphism in these rocks. If this is correct, it implies that the Eikefjord rocks were not fully subducted during the subduction of the Lower Plate/WGR.

An apparent paradox:
was the Upper Plate, that once existed between Hornelen and Håsteinen, both stationary and westward-moving "at the same time"?

In Chapter 3, it was concluded that the mylonitic Eikefjord Group in the study area is part of the Nordfjord–Sogn Deatchment Zone. In addition, the Eikefjord Group (and Lykkjebø Group) is seen to continue northwards to the southern margin of the Hornelen Devonian Massif. The rocks of both groups appear to be mylonitic in the whole area between the Håsteinen and Hornelen Massifs (Sect. 3.4), i.e. between the Sunnar Fault and the Haukå Fault (**Fig. 2.6**). In the present thesis, the whole area between the faults is interpreted as part of the Nordfjord–Sogn Detachment Zone, an interpretation also advocated by Wilks & Cuthbert (1994) after detailed field work in the area. Recently, this interpretation has slo been supported by Johnston et al. (2007b). The *implications* of this interpretation, in relation to sedimentation in the Hornelen Devonian Massif, is discussed in the following.

In the Grøndalen area, located within the southern marginal part of the Hornelen Devonian Massif, an E-W trending valley has been developed along the E-W trending axial trace of the Grøndalen syncline. The valley may be termed the "Grøndalen syncline valley". Parallel to this valley, but **1–1.5 km** further south, the Haukå Fault defines the southern margin of the Hornelen deposits. Within the "Grøndalen syncline valley", in the area east of the Grøndalen farm, the valley has mountainsides with steep to moderate slopes towards the valley bottom, which follows along the axial trace. If one stands on the north side of the valley and looks over to the southern mountainside, one sees the characteristic eastward-dipping Hornelen beds (Steel 1988) in the mountainside. Here, each bed consists of sandstones to the east, a conglomerate body in the middel, and sandstones to the west. The conglomeratic bodies, one in each bed, are marginal alluvial fans that were deposite from the southern margin, and that reached into the sandy, more distal parts of the basin. Still looking southwards, but now looking gradually upwards/eastwards in the stratigraphy, one sees that each conglomerate body is deposited somewhat eastward compared to the one below. Steel (1988) has described this ***lateral off-set stacking*** of Devonian conglomeratic alluvial fans which were buildt out northwest-wards from a *fixed point* of sediment supply to the south. The sediments were supplied from elevated source ares. Each conglomerate fan was deposited subhorisontally at a depocentre located above the listric, westward-dipping, basin-controlling fault – a fault which was probably the Nordfjord–Sogn Detachment Zone (NSDZ) reaching to the daylaght surface. As the Hornelen basin moved westwards on the subjacent NSDZ mylonites, the beds deposited above the listric fault were successively rotated to achieve the characteristic eastward dip (Steel 1988) ("roll over"-

style geometry at half-graben). *However, and this is the paradox*: since the whole area presently exposed between the Håsteinen and the Hornelen Devonian Massifs is interpreted to be an integral part of the Nordfjord–Sogn Detachment Zone, this would mean that the *source area* for the described conglomerate fans in the Hornelen basin would also move westwards on top of the detachment mylonites, and the *lateral off-set stacking* of the fans would be very difficult to explain. If the source area was indeed locked in a fixed positions relative to the subjacent WGR, and relative to the westward-moving Hornelen basin, the source area was of Lower Plate type. This would imply that the NSDZ was not present to the south of the Hornelen basin, a situation which would be contrary to the presently obsevable mylonites. If, on the other hand, the source area did actually move westwards relative to the WGR, the source area would be of Upper Plate type, resting on the NSDZ. This would be consistent with the observed NSDZ mylonites, but would, as explained above, impose difficulties on the possibility to form the *lateral off-set stacking* of the conglomerate fans.

The problem may be solved in the following way: since the Eikefjord Group mylonites represents the detachment zone, it is possible that a "time-differensiated" extension was taking place in the region. This kind of extension would mean that the Hornelen Devonian Massif developed first, with the laterally off-set stacked fans being fed from a stationary Lower Plate to the south, and that the HDM developed later, with onset of extension for the area between the Håsteinen and Hornelen Devonian Massifs, making the area an Upper Plate. It is also theoretically possible that the westward movement of the Hornelen basin – as well as the southern source-areas – occurred simultaneously, but that the Hornelen basin moved at a higher rate than the areas to the south, although this can only be speculations. In conclusion, differential movements may explain the juxtaposition of the *latrally offset-stacked* Hornelen marginal fans and the Nordfjord–Sogn Detachment Zone.

Apparent absence of NSDZ-type mylonites in the Høydalsfjorden Complex

The present section discusses the possible presence of detachment-mylonites of NSDZ-type in the lower part of the present-day Høydalsfjorden Complex. It is here of interest to review information on the tectonostratigraphic relationship between the Eikefjord Group (EG) and the Høydalsfjorden Complex (HC), and such information may be obtained by a comparison with the Dalsfjord Nappe, situated below and west of the Kvamshesten Devonian Massif. In the tectonostratigraphy of the Dalsfjord Nappe (**Fig. 2.4**, **Table 2.2.**), the Solund–Stavfjorden Ophiolite Complex (S–SOC) (Furnes et al. 1990) and the Sunnfjord Melange (Alsaker & Furnes 1994) rest tectonically on the Herland Group, which lies unconformably on the Høyvik Group, which again lies unconformably on the Dalsfjord Suite (Brekke & Solberg 1987; Andersen et al. 1990). Several correlartions have been made: the S–SOC and the Sunnfjord Melange have been correlated with the Høydalsfjorden Complex (see Ch. 4) (Furnes et al. 1990). The Høyvik Group and the Dalsfjord Suite have been correlated with the Lykkjebø Group and the Eikefjord Group (see Ch. 3) (Fig. 1 in Osmundsen & Andersen 2001; 1994). Accordingly, it is reasonable to assume that, prior to the development of the NSDZ, the Høydalsfjorden Complex was situated tectonostratigrafically on top of the rocks of the Eikefjord and Lykkjebø Groups, or similar rocks, as also suggested by Johnston et al. (2007b). Since the "anorthosite-Jotun kindred" rocks in the Eikefjord Group have been strongly involved in the detachment zone, it would be reasonable to expect the overlying Høydalsfjorden Complex to be affected.

As previously described (Sect. 4.3.7), top-to-the-west, extensional **D_3-**mylonites are present in the Høydalsfjorden Complex. These results were obtained from thin-section studies of the ultramylonites in the Høydalsfjorden Complex (HC) at the Sunnarviken area just to the south of the Sunnar Fault, i.e. at the structurally lowestmost parts of the Høydalsfjorden Complex (HC), located adjacent to the Eikefjord Group. It is, however, possible that this HC **D_3-**mylonite fabric was formed *prior* to the formation of the **D_2-**mylonites of the Eikefjord Group, since the HC mylonites are cut by the sub-Devonian unconformity and therefore appear to have formed during the **Mode-I** movements, i.e. during the westward movements of the entire orogenic wedge. The lack of NSDZ-type extensional structures in the lower parts of the HC may also be explained by large vertical displacements along the Sunnar Fault, so that the lower part of the HC, which was possibly involved in the detachment, has been dropped far beneath the present ground surface. The amount of displacement on the Sunnar Fault is at present unknown.

In conclusion, the top-to-the-west, extensional **D_3-**mylonites of the Høydalsfjorden Complex at Sunnarviken and elsewhere, is probably not related to the NSDZ. The base of the HC, however, which has probably been faulted to a level far below the present ground surface, may contain shear fabrics related to the movements that produced the **D_2-**mylonites of the Eikefjord Group.

Depth from the Upper Plate to the NSDZ in the study area, as revealed by brittle faults along the "Florø peninsula" horst

It has been mentioned above (Sect. 6.4.4) that the Lower Plate *eclogitic gneisses* that are structurally situated below the Nordfjord–Sogn Detachment Zone (NSDZ), reappears at the daylight surface near Florø town, west of the study area. Such eclogites have been observed on the island *Kinn* (**27 km** W of the study area), and on the neighbouring island *Reksta* to the east of Kinn (**20 km** W of the study area) (Kolderup 1928) (**Fig. 2.3** and **2.4**), both of which are located **12 km** and **7 km**, respectively, to the west of Florø town. Eclogites have also been found on the island of *Florelandet*, which hosts the town of Florø (see Fig. 1 in Andersen & Jamtveit 1990 and Fig. 1 in Svensson & Andersen 1991), and this occurrence lies **15 km** west of the study area. The Kinn and Reksta islands are situated on the straight westward continuation of the Florø penninssula.[(1)] The eastern areas of this peninsula, i.e. in the direction towards the village of Eikefjord, is defined by NSDZ mylonites developed in the rocks of the Eikefjord and Lykkjebø Groups, whereas in the western areas, the rocks change to more homogeneous gneisses of the Lower Plate (Michelsen 1986). Hence, when we move from E to W along the peninsula, we observe a cross-section going down throught the

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[(1)] Note: The name "Florø peninsula" has been invented to simplify naming in the present thesis. The "peninsula" actually contains the two islands Florelandet and Brandsøya, but since these islands are separated from each other — and from the proper mainland peninsula — with only a few metres of sea water, they are all grouped under the name "Florø peninsula").

Nordfjord–Sogn Detachment Zone and down into the eclogitic Lower Plate. In addition to the eclogite occurrences on the Kinn, Reksta and Florelandet islands, also smaller islands between the larger ones are likely to contain eclogites.

Eclogites are not found on the nearby islands located to the north and south of the Florø peninsula. Instead, the islands to the north consist of local occurrences of Devonian Hornelen sediments that rest on Upper Plate rocks such as muscovite gneiss, greenschists and amphibolites (Lutro 1991; Lutro & Bryhni 2000). These Upper Plate rocks may represent part of a “gneissic” Caledonian nappe of same type as the gneisses described from the Upper Plate at the Bremanger area (Sect. 2.6.2.1, and 2.7.1.5) (Hartz et al. 1994; Cuthbert 1991). Together, the Hornelen sediments and the substrate rocks in these western areas can be said to constitute a “Hornelen Upper Plate”. The islands to the south of the peninsula consist of Upper Plate rocks of Caledonian ophiolite cover-sequence and/or melange type (Lutro 1991; Lutro & Bryhni 2000), that may be incorporated into the Høydalsfjorden Complex (HC) which extends west of the study area of the present thesis. The Håsteinen sediments and its substrate of HC rocks can be said to form a “Håsteinen Upper Plate”.

The Florø peninsula displays NSDZ mylonites in the eastern part and eclogitic gneisses to the west. This penninsula, and the eclogite-containing islands to the west, form a “land array” that is **30 km** long and **2–3 km** wide in map plane, bordered on the north and south side by **1–2 km** wide fjords. The fjord on the northern side follows the westward trend of the Haukå Fault, and the fjord on the southern side the westward trend of the combined Eikefjord Fault/Sunnar Fault. North of the northern fjord and south of the southern fjord the islands of Upper Plate rocks described above, are present. This map picture suggests that the Florø peninsula and the islands to the west form a horst structure that is bordered by the above mentioned W-E oriented faults along the northern and southern sides, possibly with **km-scale** dip-slip displacements. Although the amount of displacement cannot be precisely estimated, the rock architecture may allow some rough estimates. On top of the eclogites of the western islands (Kinn, Reksta, Florelandet), the NSDZ and the Upper Plate are now missing, and if we attempt to suggest thicknesses on these missing units by comparing with areas to the north and south, we can arrive at some estimates. If we assume, for example, that the Upper Plate rocks that are present south and north of the Kinn island, have a thickness of minimum **1 km**, and that the NSDZ mylonites underlying the Upper Plates there have a thickness of minimum **1 km**, the depth to the Lower Plate eclogites, on the north and south side of the penninsula, will be minimum **2 km**. This will then also be the minimum dip-slip on the two faults. However, a thickness of **1 km** for the NSDZ is a very conservative estimate. To the south of the Håsteinen Devonian Massif, Andersen & Jamtveit (1990) estimated the thickness of the Standalen segment of the NSDZ to be **2–3 km**, and other authors have suggested similar NSDZ thicknesses from other places along the NSDZ. If we assume a similar NSDZ thickness north and south of the Florø peninsula, and add an Upper Plate thickess which we now allow to be for example **2 km**, this would mean that the depth to the eclogites on the north and south side of the Kinn island could be **4–5 km**, which would again, by implication, be the minimum dip-slip fault displacement. Hense, it seems reasonable to estimate the vertical displacement to be in the order of **2–4 km**.

Along strike eastward from the Kinn island, both the Haukå and Eikefjord Faults appear to terminate at a point about **60 km** east of the island, whereas the Sunnar Fault, branching from the Eikefjord

Fault, ends at a position about **45 km** to the east, at the point where it meets the Standal Fault. The faults are brittle in character.

The horst structure of the Florø peninsula clearly suggests that the eclogite-containing Lower Plate is actually situated below the Devonian sediment and substrate package of both the “Håsteinen Upper Plate” and the “Hornelen Upper Plate”, and that this is also the case in the utmost western areas of these plates. The horst also shows that the *thickness* of the “Hornelen” and “Håsteinen Upper Plates”, i.e. the *depth* from the daylight surface to the NSDZ, is probably never more than a few kilometres even in the westernmost parts, since larger thicknesses would, when adding the NSDZ thickness, imply extreme dip-slip values on the faults bordering the Florø peninsula.

The above reasoning may have relevance to the Sunnar fault, which in the study area separates the Eikefjord Group to the north from the Høydalsfjorden Complex to the south. It is possible that the Sunnar Fault was the locus of large dip-slip displacements, although it is uncertain whether the magnitude was equivqlent to the faults along the Florø peninsula. Large normal displacements on the Sunnar fault, bringing the south side (Høydalsfjorden Complex) downwards relative to the north side (Eikefjord Group), could explain the apparent absence of NSDZ type extensional structures in the Høydalsfjorden Complex.

To the south of the Håsteinen area, the brittle Standalen *fault* follows along, and truncates, the thick mylonites of the Standalen segments of the NSDZ. The four faults discussed here, i.e. the Haukå, Eikefjord, Sunnar and Standal Faults, are all trending roughly E-W and appear to have normal displacements. Braathen (1999) suggested that the faults formed due to N-S extension in the Late Jurassic–Early Cretaceous. Normal displacements on the Sunnar Fault to the north of Håsteinen, and the Standal Fault to the south, implies that the entire area between them, the “Håsteinen Upper Plate”, appears to be a down-faulted mega-block.

6.4.5 FOLDING OF DEVONIAN AGE

The following sections discusses the effects of Devonian folding on the Håsteinen Devonian sediments (Sect. 6.4.5.1), the Høydalsfjorden Complex (Sect. 5.4.5.2) and the Eikefjord Group (Sect. 5.4.5.3).

6.4.5.1 FOLDING OF THE DEVONIAN SEDIMENTS

General

The folding of the Devonian sediments has been explained by two strongly contradicting models; the detachment model and the "Solundian Orogeny"-model (see Sect. 2.8 and 2.9). In the detachment model, the folding is, in gross, assumed to have occurred *during* the overall post-orogenic extensional movements which also formed the Devonian basins (e.g. Osmundsen & Andersen 2001; Krabbendam & Dewey 1998). In the "Solundian Orogeny"-model, the folding is assumed to have occurred *subsequent* to and independent of the basin formation, as a result of a late phase of continued Caledonian orogenic-type contraction (Sturt & Braathen 2001). In both models, the folding of the Devonian sediments is believed to have occurred during Mid–Late

Devonian times. A *maximum* age of folding is given by the fossils in the Hornelen Devonian Massif. The fossils have been assigned to the Middle Devonan, which started at **391 Ma** according to Gradstein & Ogg (1996) or **398 Ma** according to Gradstein et al. (2004). A *minimum* age of folding is provided by palaeomagnetic datings obtained by Torsvik et al. (1988; 1996). However, the interpretation of Torsvik et al. (1988; 1996) that the palaeomagnetic data indicate folding in the **Late Devonian–Early Carboniferous** has been disputed in the present thesis (see Sect. 2.8.5.6) where it is argued that the data rather allows for folding to have occurred at *anytime* during the full length of the Devonian period, and that the folding did *not* extend, to any noticable degree, into the Carboniferous. Hence, the minimum age of folding is here suggested to be the **Devonian–Carboniferous boundary**, which corresponds to **354 Ma** according to Gradstein & Ogg (1996) or **359 Ma** according to Gradstein et al. (2004).

For reasons given above (Sect. 2.9 and 6.4.1), the explanation given by the "Solundian Orogeny"-model, for the Devonian folding, is considered less likely than the explanation provided by the detachment model. Therefore, the following section will concentrate on the explanation offered by the detachment model.

Origin of the compressional forces

The formation of the basins has been discussed earlier (Sect. 6.4.2), and was found to be a result of sedimentation occuring in depressions formed due to the crustal-scale extension. In western Norway, such a crustal-scale extension probably took place after the termination of the Scandian orogenic phase of the Caledonian Orogeny (Fossen 1992; 2000; Milnes et al. 1997; Fossen & Dunlap 1998). Since the folding of the Devonian sediments apparently occurred during the extensional movements that also created the basins, the challenge, in terms of modelling, is to find a mechanism that can produce W-E trending folds during these overall westward directed movements of the Upper Plate**.**

Fold model 1: folding due to transpression along coverging lateral ramps. (“Lateral ramp” model).

Lateral or oblique ramps may have been part of the geometry of the Nordfjord–Sogn Detachment Zone (NSDZ), as evidenced by the several E-W striking detachment segments with steeply dipping mylonite foliation, now present along the northern and southern margins of the Devonian massifs. The steeply dipping Standalen segment of the NSDZ, south of the Høydalsfjorden and Håsteinen rocks, may be an example of such a lateral ramp. In the present model, on transpression along lateral ramps (see **Fig. 6.6**), it is assumed that the E-W striking ramps are themselves not a result of the contraction that folded the Devonian sediments. Instead, the ramps are considered as primary corrugations of the NSDZ (Norton 1987; Chauvet & Seranne 1994). The fact that the E-W striking segments of the Nordfjord–Sogn Detachment Zone appear to be dipping in the direction *below* the Devonian massifs, may reflect the pimary shape of an irregular detachment zone.

From detailed studies of the folded bedding that is present along the northern margin of the Hornelen Devonian Massif, Larsen (2002b) concluded that the folding occurred during the westward movement of the basin on the NSDZ. The folding was interpreted to be a result of dextral transpression towards the basin-controlling marginal faults, as also tentatively suggested by Norton (1987). In the *Hornelen* type of setting, the marginal faults on the northern, eastern and southern sides of the basin were separating the westward moving Upper Plate from the Lower Plate. As the Upper Plate moved westwards and the basin formed in the half-graben to the east, new Devonian beds were deposited against the NSDZ itself (**Fig. 5.86**). Along the northern (and possibly southern) fault margin of Hornelen, the contact between the Devonian beds and the marginal fault caused the tranpressional shear to affect the Devonian beds directly.

For the *Håsteinen* basin, the situation was different, since the Devonian beds were not deposited with a tectonic contact against the NSDZ or marginal faults, but with a primary unconformity on top of the Upper Plate (**Fig. 6.6, 5.86**). Also, the basin did not form in a half-graben, as were the case for Hornelen, but in a ramp syncline, where the beds were deposited against a westward-dipping Upper Plate that was moving on the westward-dipping ramp (**Fig. 6.1, 6.3**). Hence, if a model on "transpression during extension" is to be used to explain the folding of the Håsteinen beds, the model would imply folding also of the *Upper Plate*. Such a model is illustrated in **Fig. 6.6**, and the model works in the following way: generally, the Devonian beds of Håsteinen have been deposited on the Upper Plate, in the ramp syncline (**Fig. 6.1**). The easternmost part of the basin, where the beds aquire their eastward dip, is not shown in **Fig. 6.6**, but can be envisaged to exist "east" of **Fig. 6.6.a**. In **Fig. 6.6**, the Devonian basin is developed in a depression formed on the Upper Plate, and the northern and southern basin margins have basinward dips. The dip of these margins is, in turn, reflecting the dip of the two lateral ramps in the subjacent Nordfjord–Sogn Detachment Zone (NSDZ). During sedimentation, all the latest deposited beds will initially span the full N-S width of the basin. As the entire basin moves westwards, the beds are transported downwards to larger depths. The basin becomes gradually narrower towards depth, since the basin margins to the S and N, have basinward dips. Conseuqently, the beds must fold, and in **Fig. 6.6**, this folding has produced the steeply dipping bed-segments just on and above the sloping basin margins. However, since the folding that results from the narrowing basin actually represents a fold model in itself, it will not be further treated here, but instead discussed separately below (as "Fold model 3"). An important element of **Fig. 6.6** is the slightly convergent strike directions of the two lateral ramps, which means that the width of the depression becomes constantly narrower towards the west. During westward movement of the Upper Plate, the Upper Plate areas located between the two ramps are therefore subjected to gradually increased contraction. In **Fig. 6.6**, this increased contraction is indicated by thickening and folding of the Upper Plate, and thereby also by folding of the Devonian beds. However, although the model may work in theory, the model has some challenging features that are worth considering.

One such feature has to do with the scale of the model. In the present thesis, it has generally been assumed that the Devonian basins had a vertical depth of **~10 km**, and that the Høydalsfjorden Complex (Upper Plate) also had a thickness of **~10 km**. This is reflected in **Fig. 6.6.a**, where the thicknesses of the two units are drawn to be equal. The thickness of the NSDZ is assumed to be **~3 km**, which is also shown with right proportions in the figure. However, as seen in e.g. **Fig. 6.6.a**, the *width* of the basin is out of proportions with the other dimensions. This has been done to ease the recognition of the geometries. For instance, at the eastern end of **Fig. 6.6.a**, the bedding-width at the basin bottom is **6 times** greater than the **~10 km** thickness of the basin, implying that the bedding-width is drawn to be **~60 km**. In comparison, the *present-day* width of the Håsteinen Devonian Massif is actually not more than **~5 km**. If we compare the present-day Håsteinen *width* of **~5 km**, with the assumed original *depth* of **~10 km** for the basin, we see that the present-day *width* is merely half of the assumed original *depth* of the basin – or vice-versa, that the original depth was twice the present width. The problem with the present-day limited width of the Håsteinen massif can be reduced if we assume that the Håsteinen sediments originally covered the entire Høydalsfjorden Complex in the area between Solheimsfjorden/Eikefjorden and Standal, a situation which is higly likely. This distance is **~10 km** in the N-S direction, implying that the basin width would then *equal* the basin depth. Furthermore, the present Håsteinen deposits represent the bottom of the basin. It is likely that the original basin margins to the south and north of Håsteinen had basinward dip directions, and this implies that the width of the basin were significantly larger at the surface than at the bottom. In the present context, the dip angle of the basin margin would reflect the dip of the lateral ramp of the NSDZ. To the south of the study area, the Standalen segment of the NSDZ has a present-day dip of around **60°** (Bryhni & Lutro 1991a, 1991b, 2000a, 2000b). If we assume that the *original* basin margins to the north and south were also dipping at an angle of **60°**, at the time of ORS deposition, then the *surface* of the basin would in fact be **6 km** wider on the northern side, and equally wider on the southern side, yielding a total basin width of (10 + 6 + 6 =) **22 km** at the surface. It is worth noting that an increased basin width northwards would possibly make the Håsteinen basin interfere with the Hornelen basin. This is due to the short distance between the Høydalsfjorden Complex and the Hornelen sediments. At present, the shortest lateral distance from the Høydalsfjorden rocks – once covered with Håsteinen sediments – to the Hornelen deposits, is around **3 km**, as measured from Sandviken to Haukå (**Fig. 2.3**). Hence, if the surface of the original Håsteinen basin did actually reach **6 km** north of its assumed “basin bottom” at Sandviken, the Håsteinen basin would appear to coalesce with the Hornelen basin. However, as argued in Sect. 6.4.4, it is not obvious that the two basins, during deposition, had the same *relative* positions as today. In summary, the present narrow width of **5 km** for the Håsteinen basin need not be an objection to the fold model.

A second issue concerns the possibility that two E-W trending zones in the *Upper Plate*, located along the two lateral ramps, but outside the depression, will be subjected to N-S directed *extension*. This is due to the fact that orientations of movement vectors on the depression (basin) side of the lateral ramp, differ from the vector orientations present outside the depression. In the model, the movement vectors outside the depression indicate a general westward movement of the Upper Plate. On the depression side of the ramps, however, the movement is parallel to the strike direction of the ramps, which is somewhat oblique to the movement direction outside. If the part of the Upper Plate that is located between the ramps in **Fig. 6.6.a**, *continues* to be located between the ramps during the westward movement of the Upper Plate/Devonian sediments, then the Upper Plate in the depression will gradually contract and, hense, gradually “pull back” from the Upper Plate outside. This will form a zone of extension along each of the ramps.

A third aspect of the model concerns the question of whether the contraction of the Upper Plate and the Devonian beds would actually lead to folding, or whether it would lead to thickening without folding. Buckling-type folding is a deformation style that depends on the existence of layers with competence contrasts. On a basin scale, the Håsteinen deposits consist of monotonous conglomerates with no such layers, and it may therefore be suspected that the Håsteinen unit would not deform by buckling, but rather by uniformal thickening. The rocks of the Høydalsfjorden Complex, on the other hand, contain layers that had been folded into E-W trending folds prior to the formation of the Devonian basin, and these outcrop-scale folds could have been in the posisiton of being tightened during Devonian contraction. The Høydalsfjorden layers that are folded are defined by bedding in turbiditic greywackes (locally greenwackes), and less pronounced bedding in micaschists and local metapsammites – and the bed thickness is usually **0–20 cm**, with a maximum around **50 cm**. However, when the rocks are considered on a hundredmeter or kilometere scale, they appear to form a very monotonous grey or green unit of metasediments. This monotony reflects the rapid sedimentation in a back-arc basin, where the sedimentation gave no pronounced layering of different rock types that can be traced through the unit. When such a monotonous unit is subjected to the N-S contraction, it is possible that the unit would respond by uniform thickening rather than large-scale folding. It appears that, if the Høydalsfjorden Complex responded by thickening, so would the “unlayered” Håsteinen deposits.

A fourth element needing attention concerns the orientations of the movement vectors located between the converging ramps. In the model (**Fig. 6.6**), the movement vectors between the ramps must be oriented parallel to the strike directions of the ramp, to generate the contraction resulting in folding. However, it may be questioned whether the latral ramps of the NSDZ would actually be able to force the initial Upper Plate volume to continue to stay between the converging ramps during the westward movement of the plate. A different development may be envisaged, where the Upper Plate would instead “climb” out of the narrowing depression during the westward movement, hence following a movement direction parallel to the movement vector outside the depression. The extent to which this would happen would depend on the relationship between the contractional forces generated by the slope of the ramp, and the ability of the rocks to resist these forces.

A fifth subject regards the question of whether the *strike* orientation of the *present-day* lateral ramps located between Nordfjord and Sogn, will actually be consistent with the model (**Fig. 6.6**). All the Devonian deposits in this area have lateral ramps (segments of the NSDZ) along their northern and southern margins, making each deposit confined by a ramp pair. As seen in **Fig. 2.3**, it appears that the strike orientations of several of the present ramp pairs are actually *diverging* towards west, instead of converging as shown in the model. This is particularly the case with the lateral ramps around the Solund and Kvamshesten deposits. Also the Hornelen massif has slightly diverging ramps. If these diverging strike directions were also present during the westward movement of the Upper Plate, it would be difficult to apply the model to these basins. Regarding the Håsteinen deposits, the ramp along the southern margin, i.e. the Standalen segment of the NSDZ, swings – in the westward direction – into parallelism with the southern margin of Hornelen massif, or even into a a slightly converging angle. The fold model may therefore be relevant in the Håsteinen case.

In summary, the lateral ramp model with folding due to lateral transpression may work, although it seems to have problematic sides.

Fold model 2: folding due to "ridge-shaped" frontal ramp. ("Ridge-ramp" model).

In this model (**Fig. 6.7**), the "frontal" ramp of the detachment defines a nonplanar surface that has a ridge-like shape, with the ridge-crest plunging down the ramp. In the model, this ramp geometry is in contrast to the *lower flat* (below the ramp) which is planar (**Fig. 6.7**). The model works as follows: sedimentation will take place in the ramp basin as illustrated in **Fig. 6.1** and **6.3**. Beds are first deposited horizontally (**Fig. 6.7**), with a depositional unconformity against the dipping Upper Plate, and are then transported down the ramp, along with the Upper Plate. At the base of the ramp, the Devonian beds will have aquired an eastward dip which mirror the dip of the ramp (see **Fig. 6.1** and **6.3**). The unique feature of the model in **Fig. 6.7** is the combination of the non-planar, ridge-shaped ramp, and the planar lower flat. This combination will, as a geometrical necessity, lead the Devonian beds to form a fold. Hense, no N-S directed contractional forces are present that can bend the former planar bed. Instead, the fold is formed directly during the downward transport of the beds, purely as a result of the geometrical relationships. This is better understood by comparing with **Fig. 6.1** and **6.3**. In these figures, it is seen that the dip towards east is aquired in one particular part of the profile: the part just above the ramp. Hense, the following geometrical relationships exist: if the ramp is planar, the Devonian bed will be planar. If, however, the ramp has a ridge-shape, then the beds aquiring their dip there, will also be ridge-shaped, and this produces beds with a fold-like shape.

The attractive feature of the model is the possibility to create a fold structure *without* applying a N-S directed contraction to impose mechanical bending of a preceding planar bed. Instead the fold is formed purely as a result of the combination of extension and the ramp shape. The model may explain the fold geometry of the beds in the Håsteinen Devonian Massif, i.e. the Osstrupen syncline.

Fold model 3: folding due to narrowing basin at depth. ("Narrowing basin depth" model):

In this fold model (**Fig. 6.8**), the sedimentary beds are experiencing the effects of a basin that has a gradually narrowing width towards its depth. As mentioned above ("Fold model 1"), the basis for the present model is the fact that most basins have margins inclined in a basinward direction. Furthermore, the basin would have to form in a depression between two lateral ramps, as in Fold Model 1, but in the present case, the ramps do not need to have convergent strike directions. The Devonian basins of western Norway probably had steep to moderately dipping margins on their north and south sides, as well on the eastern side. In the Håsteinen basin, the sediments accumulated above a westward-dipping ramp surface, that in the present model is envisaged to be a more planar surface. With such a basinward dip of the southern and northern margins, the width of the basin had to be significantly wider at the surface than in the deeper parts. In a **10 km** deep basin with margins dipping **60°**, the surface width is **6 km** wider at the surface than at the bottom. Hence, the downward movement of a horisontal layer that was deposited against the W-dipping Upper Plate, would cause the layers to fold passively (**Fig. 6.8**).

A similar model of passive folding has been proposed by Norton (1987), who used it to explain the folding in the Kvamshesten Devonian Massif. Seranne et al. (1989) used a similar model to explain folding of all the Devonian basins of western Norway. The fold model succseeds in explaining the folding present in the Håsteinen deposits.

Brief evaluation of the three fold models.

Fold Model 3, “folding due to narrowing basin at depth” (“narrowing basin depth” model, **Fig. 6.8**) appears to offer the simplest explanation. It seems likely that such space problems will occur, leading to the folding. This model is closely connected to Fold Model 1, “folding due to transpression along converging lateral ramps” (“Lateral ramp” model **Fig. 6.6**), but in Model 3, converging strike directions for the lateral ramps are not needed to create the folding. A non-planar ramp as in Model 2, “folding due to ridge-shaped frontal ramps” (“Ridge-ramp” model **Fig. 6.7**), may have contributed. However, the presence of a very pronounced ridge-ramp morphology combined with reduced basin-width at depth could lead to “double”-folding, i.e. one fold from each process, but this is not observed in the study area. Since the basin anyway had a limited N-S width compared to its depth, and the northern and southern margins must have dipped to the south and north respectively, it appears that the reduced basin width at depth would be sufficient to create the fold. If so, the contribution from the ridge-ramp model is not needed to explain the folding. In summary, it seems most likely that the folding is a result of the reduction of basin width towards depth (Fold model 3).

Discussion of folding mechanism

The folding of the sediments in the HDM may have occurred during varying rheological conditions, and may theoretically have been of the following types: (1) soft sediment deformation; (2) brittle deformation and (3) semiductile/ductile deformation. Before dealing with these alternatives, the term “syn-sedimentary” will be discussed, as the term may have a wider meaning than “soft-sedimantary”.

Explanation of the term "syn-sedimentary deformation". Since both the *sedimentation* of the Håsteinen basin – as well as its *deformation* (that produced folding and axial planar cleavage in the basin) – occurred during the same extensional movements, the cleavage-forming deformation can be characterised as "syn-sedimentary". The connection is as follows: if it is correct that the basin received sediments that reached a true vertical thickness of **~10 km**, which seems reasonable since the temperature in the HDM has been **300 +/- 50°C** (and since a normal thermal gradient of **30°C/km** apperars likely), it means that the term *"syn-sedimentary deformation*" may have two different meanings: (1) *Soft-sediment deformation* of unconsolidated sediments: here the deformation would occur at or near the surface during ongoing sedimentation. This deformation would be "syn-sedimentary deformation" *sensu stricto*, and is hereafter termed "soft-sedimentary" deformation. (2) *Ductile (plastic), semi-ductile or brittle deformation* of lithified sediments: at the same time as the sedimentation was still taking place at the basin surface, the consolidated sediments at larger depths (down to **~10 km**) could be experiencing ductile (plastic), semi-ductile or brittle deformation. Such deformation of coherent rocks would be taking place at a time when the basin was still developing as a result of the extensional movements that controls the basin formation. Hence, since the sedimentation was still going on at the basin surface, the deformation in the coherent rocks at depth would also be "syn-sedimentary" in a basinal context. The deformation of both the unconsolidated sediments and the coherent rocks occurred during the same continuous dynamic prosess, and a similar development presumably took place in all the Devonian basins of western Norway.

The HDM has been quite strongly folded compared to the other Norwegian Devonian basin, as is documented by the interlimb angle of **103°** for the Osstrupen syncline. Yet, the conglomeratic clasts throughout the HDM show no visually detectable signs of strain (e.g. fracturing, or ductile/plastic flattening or constriction). If the folding of the HDM took place after the rocks were buried to **300 +/- 50°C** (**~10 km**), then the deformation would probably have been ductile (plastic) or semi-ductile, and reorganisation of the clasts by particulate flow to accommodate the strain would be less likely. The flexural flow that would presumably have taken place in the fold limbs would not necessarily – despite the fairly large amount of folding present in the HDM – produce a visually detectable distortion of the clasts. Also, rotation of individual clasts in the semiductile/ductile state to adjust to the strain cannot be completely excluded. Yet, the apparent absence of visually observable distortion could suggest that some of the folding in the HDM occurred prior to complete consolidation of the sediments. Nevertheless, the axial planar cleavage present in the parasitic folds at the Gravaneset sandstone unit show that most of the shortening of the HDM must have occurred after burial to semiductile/ductile conditions. The amount of possible soft-sedimentary deformation is generally uncertain, but is assumed to have been limited The role and amount of possible brittle fractures, forming in the time period between the consolidation/cementation of the sediments, and the burial to semiductile/ductile conditions, is at present unknown. A separation of such possible Devonian brittle faults from the later faults (Permian, Mesozoic, etc.) would not be easy, although a possible systematic orientation related to the Devonian strain field could provide a possible way of identification.

In conclusion, the presence of axial planar cleavage in the Gravaneset sandstone unit is a clear evidence that most of the deformation of the HDM occurred in the *semiductile/ductile* (plastic) state. Throughout the Håsteinen deposits, indications of soft sedimentary deformation have not been found. Although some degree of strain-related clast-reconstitution in the unconsolidated state is possible – since clasts in the HDM generally lack distortion – such soft-sedimentary deformation is considered to be unimportant during the formation of the Osstrupen syncline.

6.4.5.2 EFFECTS OF DEVONIAN FOLDING ON THE EIKEFJORD GROUP

The Devonian post-Scandian deformation has effected the rocks of the Eikefjord Group in two ways; firstly, by the development of the extensional $\mathbf{S_2}$-mylonites, which makes the rocks part of the Nordfjord–Sogn Detachment Zone (discussed in Sections 3.4, 4.3.5.3, and 6.4.2.1), and secondly, by the $\mathbf{F_3}$-folding which will be considered here. (The block diagram of **Fig. 6.9** gives an overview of the main structural feartures of the study area).

Two questions are of particular interest concerning the $\mathbf{F_3}$-folds in the Eikefjord Group: (1) did the Eikefjord Group $\mathbf{F_3}$-folds form at the same time and by the same process as the Osstrupen $\mathbf{F_1}$-fold in the HDM?; and (2) did this folding occur during or after the extensional movements? These questions are discussed below.

Coeval formation of the F_3-folds of the Eikefjord Group and the F_1-Osstrupen syncline of the Håsteinen Devonian Massif

In the present work, it is suggested that the $\mathbf{F_3}$-folding of the Eikefjord Group (EG) took place in the same deformational process as the $\mathbf{F_1}$-folding of the Håsteinen Devonian Massif (HDM), which produced the Osstrupen syncline. The arguments in favour of this are as follows: (1) The EG-$\mathbf{F_3}$ and HDM-$\mathbf{F_1}$-fold axes have parallel trends (**Fig. 6.9**) (Sect. 3.4 and 5.5.3.3). This is a neccesary, but not sufficient condition for stating that the folds formed by a common strain system. (2) The detachment zone has been interpreted as a feature controlling the basin formation. This interpretation is apparernt from the overall rock architecture of the area, where the eclogite facies rocks of the Lower Plate are juxtaposed with greenschist facies rocks of the Upper Plate, separated by the Nordfjord–Sogn Detachment Zone (NSDZ) with its abundant presence of shear sense indicators showing top-to-the-west sense of movement. The $\mathbf{S_2}$-mylonites of the NSDZ have been interpreted as a Mode-II fabric formed at or just before **394 Ma**, and recently Johnston et al. (2007b) suggested that the zone moved in the time interval **410–400 Ma** (Sect. 6.1.1). However, for the present discussion, the exact timing of the movements are not a crusial factor. Since a great portion of the detachment $\mathbf{S_2}$-mylonites apparently was formed prior to the $\mathbf{F_3}$-folding, and the $\mathbf{S_2}$-mylonites probably were a controlling factor for the formation of the Devonian basins, also the Devonian sediments must have been present prior to the $\mathbf{F_3}$-folding. It is even possible that the $\mathbf{S_2}$-mylonites of the Eikefjord Group could have been retrograded towards the greenschist facies, before the EG $\mathbf{F_3}$-folds developed, implying that the folding possibly occurred at an advanced stage of the extensional movements, and thereby at a stage when the Devonian sediments were obviously present. Since the Devonian sediments were present during the formation of the EG-$\mathbf{S_2}$-mylonites, the folding of the same mylonites was most likely accompanied by folding of the Devonian deposits. Thus, the ductile $\mathbf{F_3}$-folding of the Eikefjord Group may be considered as compatible with the semiductile/ductile late phase of the $\mathbf{F_1}$-folding of the HDM, that produced the Osstrupen syncline. In summary, it is therefore reasonable to assume that the deformation which produced the $\mathbf{F_1}$-Osstrupen syncline in the HDM also produced the $\mathbf{F_3}$-folds in the Eikefjord Group.

Above, the folding of the HDM into the $\mathbf{F_1}$-Osstrupen Syncline has been suggested to be caused by either (i) transpression towards lateral ramps, (ii) a ridge-shaped ramp, or (iii) reduced basin-width at depth, or combinations of the three (Sect. 6.4.2.1 and 6.4.5.1). In all these models, the folding is a result of intrabasinal forces, implying that regional external N-S contractional forces are not needed. Also in the Hornelen Devonian Massif, folding has been interpreted as a result of intrabasinal processes, notably along the northern margin of the basin, where Larsen (2002b) interpreted the folding as a result of transpression along the basin-controlling fault during westward movement of the basin. The Eikefjord Group mylonites are located in the area between the Hornelen and Håsteinen Devonian massifs. If the folding of both these basins are related to intrabasinal processes, i.e. processes connected directly to the extension, then the N-S contraction leading to the $\mathbf{F_3}$-folding of the mylonites is possibly related to contraction operating *within* the shear zone – although it cannot at the present stage be excluded that regional *external forces* contributed in the formation of the $\mathbf{F_3}$-folds.

The Eikefjord-$\mathbf{F_3}$-fold axes have trends parallel to the Osstrupen syncline $\mathbf{F_1}$-fold axis, but in addition the trends are also parallel to the Høydalsfjorden $\mathbf{F_2}$-fold axes. However, the formation of the Høydalsfjorden $\mathbf{F_2}$-fold axes cannot be the result of the process folding the Håsteinen beds, since the Høydalsfjorden $\mathbf{F_2}$-folds are cut by the sub-Devonian unconformity (**Fig. 6.9**). This is further discussed in the next section (Sect. 6.4.5.3).

Significance of parallel Eikefjord Group F_3-fold axes and L_2-lineations for the time of folding

In the Eikefjord Group, the $\mathbf{F_3}$-fold axes are parallel to the $\mathbf{L_2}$-stretching lineations, and both have a E-W to ESE-WNW trend and shallow plunges (**Fig. 6.9**). As mentioned in Sect. 3.4.3, this parallelism can be explained in two ways; either, **(1)** the $\mathbf{D_2/L_2}$-shear movements and the $\mathbf{F_3}$-folding occurred simultaneously and are genetically related, or **(2)** the $\mathbf{D_2/L_2}$-movements occurred first and the $\mathbf{F_3}$-folding later, and their parallelism is coincidential due to similar orientation of two independent strain systems.

(1) The first alternative implies that the parallelism formed because $\mathbf{F_3}$-folds formed *simultaneously* with the development of the $\mathbf{S_2/L_2}$-mylonite that defines the limbs of the $\mathbf{F_3}$-folds. If both the shear rate and the amount of movement was small, and if the movements were taken up by the abundant and laterally extensive "low friction" micaceous and chloritic lamina in the mylonite, it is possible that such $\mathbf{D_2}$-shear could take place after the formation of the $\mathbf{F_3}$-folds, without destroying the folds. However, any large amount of shearing subsequent to the formation of the $\mathbf{F_3}$-folds would probably destroy or modify the folds. Another way to make "folding during shearing" possible could be that the shear did not operate in the *entire* vertical coloumn of the mylonite package at the same time, a situation which is likely. If, for example, at some stage during shearing, the shear was localised to a lower structural level in the mylonite package, the upper part of the mylonite would effectively become part of the Upper Plate, and fold together with this unit. The $\mathbf{F_3}$-folding of this $\mathbf{S_2}$-mylonite could then occur simultaneously with the continued shearing in the mylonites further below. Hence, it is possible that the $\mathbf{L_2}$-lineations and the $\mathbf{F_3}$-folds are genetically related.

(2) In the second alternative, the parallelism is coincidential.This situation, where the $\mathbf{F_3}$-folds were formed *entirely after* the formation of the $\mathbf{S_2}$-mylonite package and $\mathbf{L_2}$-lineations, i.e. after cessation of the top-to-the-west movements on the NSDZ, would, of cource, make it problematic to explain the folding of the Eikefjord Group as a result of transpression *during* the extensional process. Hense, the folding would then need an alternative explanation. In this case, post-extensional, N-S directed compressional forces would have to be envisaged to deform the $\mathbf{S_2}$-mylonite package and produce the $\mathbf{F_3}$-folds. However, this possibility raises some problems, as it may be questioned whether such a contractional event has in fact occurred. The **Mode-I** top-to-the-west displacement of the entire orogenic wedge of South Norway, on the basal décollement zone (Fossen 1992, 2000), has been dated to **402–394 Ma** (Fossen & Dunlap 1998). These movements suggest that the Caledonian orogenic *contraction* had in fact reached its cessation prior to this time interval. The top-to-the-west movements reflect a situation where the Laurentian and Baltic plate movements had changed from convergent to divergent (Fossen & Dunlap 1998). The **Mode-II** Nordfjord–Sogn Detachment Zone, which developed at the last stages of, and continued its movements subsequent to, the **Mode-I** displacements, further emphasises that the Caledonian contraction had ended in south Norway. Therefore, a model implying a new phase of N-S contraction that must entirely ***postdate*** the top-to-the-west displacements along the **S_2**-mylonites of the Eikefjord Group (NSDZ), implies that a new phase of N-S orogenic type contraction would have to be invoked.

A model applying such a N-S contraction, where the contraction is supposed to entirely ***postdate*** the top-to-the-west movements, would accord with the "Solundian Orogeny"-model. However, as previously argued in the present thesis (Sect.6.4.1), such a new phase of N-S contraction, taking place ***subsequent*** to the extension, has not been substansiated in the literature, and is considered problematic.

In central Norway (Møre–Trøndelag), NE-SW trending folds are abundant, for example in the Trondheimsleden area (Hitra–Smøla–Ørlandet–Fosen) where Devonian ORS deposits are deformed into folds with fold axes of this trend. Several authors have suggested models where such folds are interpreted to be a result of contraction caused by sinistral shear movements between the Laurentian and Baltic plates in this region, at the last stages of the Scandian contractional orogeny (Tucker et al. 2004; Braathen et al. 2002; Osmundsen & Andersen 2001; Krabbendam & Dewey 1998). These authors also suggested that large-scale, detachment type shear sones with top-to-the-SW mylonites, were formed in the same sinistral shear system, and that these extensional shear zones controlled the formation of local Devonian Old Red Sandstone basins. Althoug the models may differ slightly, they all advocate that the formation of the NE-SW trending folds occurred simultaneously with the mylonite-producing, top-to-the-SW displacements.

The very characteristic NE-SW trend of the folds in the Møre–Trøndelag region makes this area different from the regions further north (Nordland), where the picture is dominated by large-scale detachment zones that in map view follow around Caledonian nappes such as the Helgeland Nappe Complex (see Braathen et al. 2002). The Møre–Trøndelag region also differs from the regions to the S and and SW, where the large Western Gneiss Region separates the Møre–Trøndelag district from the geological units of Western Norway, including the Nordfjord–Sogn Detachment Zone (NSDZ) and its southward continuation, the Bergen Arc Shear Zone (BASZ). With a distance of at least **230 km** from the Bergen area to the Møre–Trøndelag extensional detachment north of Stadtlandet, it is not considered very likely that a sinistral shear in the Møre–Trøndelag district would lead to folding as far south as the Bergen area (see Sect. 6.4.1), where the BASZ form a fold.

In the Basin and Range district (USA), which has been subjected to Tertiary crustal-scale extension, detachment zones form fold-like structures with fold axes oriented parallel to the shear direction of the systems. These fold-like structures have been interpreted to be primary corrugations of the detachment zones, formed *during* the extensional movements (Fletcher & Bartley 1994; Mancktelow & Pavlis 1994; Yin 1991).

In western Norway, Chauvet & Seranne (1994) analysed E-W trending folds in the Nordfjord–Sogn Detachment Zone (NSDZ), and concluded that the folds formed during the top-to-the-west movements on the zone, as a result of permutation of the maximum stress from a vertical orientation at the onset of extension to a horizontal N-S orientation in the subsequent stages. The permutation was explained by the progressive unloading of Western Norway due to the extension. Hence, the model may be seen as explaining the folding of the NSDZ mylonites as a result of forces internal to the extensional system, implying that external N-S contraction is not needed.

In conclusion, the **F_3**-folds in the Eikefjord Group were most likely formed at the same time as the Osstrupen **F_1**-syncline in the HDM, and the **F_3**-folding probably took place during the overall extensional movements. Although contributions from a minor late phase of N-S crustal shortening cannot be totally excluded, it is considered less likely.

6.4.5.3 EFFECTS OF DEVONIAN FOLDING ON THE HØYDALSFJORDEN COMPLEX

The folding of the **S_0**-bedding present in the Håsteinen Devonan Massif (**HDM**), into an **F_1**-syncline with an interlimb angle as small as **103°** (the Osstrupen syncline), is bound to have had some sort of effect on the rocks in the subjacent Høydalsfjorden Complex (**HC**), since the Devonian sediments are situated with a primary unconformity on the substrate. This is further discussed in the following.

The most prominent phase of folding documented in the Høydalsfjorden Complex produced the **F_2**-folds which have the same azimuth as – and locally overlapping plunge to – the **F_1**-fold axes of the Osstrupen syncline (**Fig. 6.9**) (Sect. 5.5.3.3). However, since it has also been demonstrated that these HC **F_2**-folds are cut by the sub-Devonian unconformity (**Fig. 6.9**) (Sect. 4.3.5.2), they cannot be a result of the N-S contraction that folded the HDM. If the folding of the HDM led to any folding in the HC rocks, this folding would have to postdate the HC **F_2**-folds. However, in the HC rocks it has not been possible to identify with certainty a post-**D_2**-fold phase which could be related to the folding of the Devonian sediments. A paradoxical situation thus appears to be present: if the Devonian N-S deformation could produce the quite prominent **F_3**-folds in the Eikefjord Group, as well as the **F_1**-Osstrupen Syncline in the HDM, the deformation should also have been able to produce some sort of deformation in the Høydalsfjorden Complex. This apparent paradoxical situation might be resolved as follows:

The overall parallel trend of HC **F_2**-fold axes/axial planes – and HDM **F_1**-fold/axial plane – means that the strain systems operating before and after deposition of the HDM were largely coaxial. It is therefore possible that folds being related to Devonian N-S contraction are actually *present* in the Høydalsfjorden Complex, but that they have not been detected since they are parallel to the already formed **F_2**-folds. The possible **F_4**-folds in the HC might be candidates for folds that were related to the N-S contraction that also folded the HDM (Sect. 4.3.8). Such **F_4**-folds have not been recorded in the field, but their existence may be inferred from the impression that the **S_3**-mylonite foliations show variable dips, whilst maintaining an overall ESE-WNW strike. In addition, it is theoretically possible that some of the substrate **F_2**-*kink folds* might actually be related to the Devonian N-S contraction. Furthermore, it is possible that the HC **F_2**-folds that are cut by the Devonian unconformity, have been somewhat “tightened” or refolded during the Devonian N-S contraction, where the degree of tightening or refolding would depend upon the original **F_2**-fold orientation at the onset of the Devonian N-S contraction. Although such tightened or refolded **F_2**-folds have not been observed in the field, their possible presence cannot be excluded.

The only place where the Devonian folding can be seen to affect the Høydalsfjorden rocks on a mesoscopic scale, is at the Gravaneset area in the far west (**Plate 4** and **5**), where the unconformity is folded.

Here, the Devonian Gravaneset Sandstone Unit rests with a depositional unconformity on turbiditic greywackes and gabbros of the Høydalsfjorden rocks. The orientation of the unconformity below the Devonian sediments are roughly parallel to the orientation of the Devonian bedding. The folding of this Devonian sandstone bedding has produced the Gravaneset **F_1**-folds (**Plate 4** and **5**), which are parasites on the large Osstrupen **F_1**-syncline. The folded Devonian bedding at Gravaveset is roughly parallel to the unconformity, and the equally folded *unconformity* is thus a result of Devonian folding. However, despite the fairly prominent folding of the unconformity, and hense the folding of the subjacent Høydalsfjorden rocks near the unconformity, the latter rocks do not appear to display particular features that can be related to the Devonian folding of the unconformity.

The Osstrupen syncline is a very simple fold structure with straight limbs. Parasitic folds are absent, apart from the very minor parasitic folds at Gravaneset in the far west. The straight bedding in the Osstrupen limbs opens for the possibility that the underlying HC rocks were, during the N-S contraction, “bent” about a subhorizontal “fold axis” that we can envisage was situated just below the HDM rocks, in a position along the axial plane of the Osstrupen syncline. In such a “bending” process, where the HC “limbs” may have been rotated in a more or less rigid manner, it is possible that only a rather small degree of outcrop-scale folding would develop in the substrate. Such a bending would be consistent with “Fold model 3” for the folding of the Håsteinen beds (“Narrowing basin depth” model, Sect. 6.4.5.1). Moreover, it has not been possible, in the Høydalsfjorden Complex, to detect metamorphic retrogression which could be related to the Devonian N-S deformation, and this may be taken to suggest that the degree of ductile deformation has been limited. The stronger degree of **F_3**-folding in the <u>Eikefjord Group</u> could then be a result of processes confined to the shear zone (NSDZ).

The Main Map (**Plate 1**) and the **Road Logs** (**Appendix A**) display structural data that show the orientation of **S_1**-foliations as well as **F_2**-fold axes and axial planes in the Høydalsfjorden Complex, and it is of interest to see whether these data yield some information on the later N-S contraction that also folded the above-lying HDM – and particularly whether the above described “bending” of the Høydalsfjorden Complex is substantiated. In the HC rocks north of the HDM, i.e. in the whole area located between the fjord Høydalsfjorden and the lake Vassetvatnet, the main map (**Plate 1**) shows that most of the recorded **S_1**-foliations and the **F_2**-axial planes have a southward dip. Better exposures of the structural features are available in the western parts of the study area, where road-sections form a N-S traverse through the HC rocks on both sides of the fold closure of the Osstrupen syncline. At a first glance, the soutward dip of **S_1**-foliations and the **F_2**-axial planes (main map, **Plate 1**) may seem to be compatible with a model on upward bending (folding) of the HC rocks in the areas north of the axial trace of the HDM **F_1**-Osstrupen syncline. However, on the **Road Logs no. 1 and 2**, which provide profiles oriented roughly N-S through the Høydalsfjorden rocks in the NW part of the study area, it is seen that the **S_1**-foliations (= limbs on **F_2**-folds) are dipping both ways, i.e. to the S and N, with neither of the dip directions being in clear overweight. Moreover, a majority of the **S_1**-foliation data on the Main Map (**Plate 1**), recorded north of Håsteinen, are located in the steep north-sloping mountainside. Hence, the recordings may have been biased by a more frequent registration of the southward-dipping foliation fabrics, since these fabrics stand out with an “easy-to-record” high angle on the rock surface, whilst the north-dipping foliations are parallel to the mountain side and therefore easier to overlook. On both the Main Map (**Plate 1**) and the **Road Logs no. 1 and 2**, the orientations of **F_2**-*axial planes* are also seen to dip mainly to the S, although the opposite N-ward dip is also present.

We now move to the Høydalsfjorden rocks located on the south side of the HDM, to consider HC data recorded on the south side of the axial trace of the Håsteinen Osstrupen syncline. In the HC rocks here, the amount of structural data on the Main Map (**Plate 1**) is very limited, and the **S_1**-foliations dip both to the N and S, although mainly to the north. The scarsity of structural data on the map may be compensated by a road log: **Road Log no 3**, which follows nearly from the Devonian/substrate contact and southwards to Store Høydal, displaying over **500 m** continuous exposures, provides large amounts of data. The **S_1**-foliation dips both to the N and S, and once again with no clear overweight to any side. However, the **F_2**-*axial planes* are still dipping largely to the south, i.e. with the same dip-direction as the one on the *north* side of the axial trace of the Osstrupen syncline. This appears to suggest that in the Høydalsfjorden rocks, the orientations of the **S_1**-foliations and **F_2**-axial planes cannot *alone* be used to infer the geometry of any HC folding related to the formation of the Osstrupen syncline. Nevertheless, the **~60°** dip of the Devonian bedding in the limbs of the Osstrupen syncline means that the subjacent Høydalsfjorden rocks must have been rotated accordingly. This means that a Devonian fold must have formed, which may be termed the “Høydalsfjorden syncline”, and which is a mirror image of the above-lying Osstrupen syncline.

In conclusion, the folding of the unconformity at Gravaneset is the only clear evidence that the Devonian N-S contraction led to mesoscopic folding of HC rocks. In addition, however, the Høydalsfjorden Complex possibly contains mesoscopic **F_4**-folds that formed during the N-S contraction that also folded the above-lying Håsteinen Devonain sediments. The apparently variable dip-orientations of the top-to-the-west HC **S_3**-mylonite foliation may be a result of such **F_4**-folding. Elsewhere in the Høydalsfjorden Complex, the separation of these folds from the earlier **F_2**-folds has been impossible due to the parallelism of the two fold generations, and possibly also due to a very modest deformation of the HC rocks during the Devonian N-S contraction. However, **F_2**-folds may have been slightly tightened or refolded. The folding of the Håsteinen rocks, which resulted in the Osstrupen syncline with limbs dipping **~60°**, must implicate a similar amount of rotation of the subjacent Høydalsfjorden rocks, forming a corresponding “Høydalsfjorden syncline”.

6.5 SUMMARY AND CONCLUSIONS

General

In the study area, three tectonostratigraphic units are present. These are from base to top, the Eikefjord Group (EG), the Høydalsfjorden Complex (HC) and the Håsteinen Devonian Massif (HDM). Each of the units displays ***different*** and complex tectonometamorphic histories. The present section (6.5), however, will summarise the geological development of the study area in *chronological* order, from the oldest to the youngest events.

Earliest geological history

The earliest geological history in the study area can be found in the Eikefjord Group, which contains Precambrian orthogneisses dated to **1511 +/- 64 Ma** (Abdel-Monem & Bryhni 1978) as the minimum intrusive age. The rocks consist of meta-anorthosite, grey monzonitic biotite-epidote gneiss, amphibolites, meta-gabbro and minor micaschist, and have been assigned to the "anorthosite-Jotun Kindred" which are correlated with the Jotun nappe and the Dalsfjord Nappe. Outside the study area, Bryhni et al. (1981) have reported granulite facies relicts of possible Precambrian origin in the Eikefjord Group. Inside the study area, however, remnants of an amphibolite facies mineralogy define the highest metamorphic grade in the rocks. The amphibolite facies metamorphism is thought to be Caledonian or post-Caledonian, as detailed below. The amphibolite facies mineral assemblage is part of the dominating and penetrativ top-to-the-west mylonite fabric of post-Caledonian age, but has been more or less retrograded to greenschist facies during the shear. All possible Precambrian or pre-mylonitic (pre-amphibolite facies) tectonometamorphic structures in the Eikefjord Group have been collectively assigned to a $\mathbf{D_1}$-phase, which is not discussed in the present thesis.

Caledonian (Scandian) contractional geological history

The Caledonian (Scandian) top-to-the-east development of the rocks in the study area can be found particularly in the rocks of the Høydalsfjorden Complex (HC). In the Eikefjord Group, the only possible Scandian features are the amphibolite facies mineralogy –that has been strongly retrograded during the post-Scandian NSDZ shear (see below). The Høydalsfjorden Complex essentially consists of meta-sediments which usually have a well developed bedding ($\mathbf{S_0}$). In certain areas, the meta-sediments are intruded by gabbros. The rocks in the complex are interpreted to represent either the cover sequence to an ophiolite – and/or a melange. The HC-rocks possibly originated in an oceanic island arc/back-arc environment, as the gabbros have intruded turbiditic meta-greywacke sequences at the oceanic stage. The occurence of large amounts of arkosic to quartzitic meta-psammites may suggest proximity to a continental source for these sediments. The rocks have been correlated with the Solund–Stavfjorden Ophiolite Complex which has been dated to **443 +/- 4 Ma**

(Dunning & Pedersen 1988), and a similar age is therefore reasonable for the formation of the Høydalsfjorden Complex. The rocks may also display similarities to a melange, and may possibly resemble the Sunnfjord Melange, which is an obduction melange located west of Kvamshesten Devonian Massif, or the Kalvåg Melange, which is a sedimentary melange located at Bremanger west of the Hornelen Devonian deposits. The Høydalsfjorden rocks were obducted onto the Baltoscandian platform during the Scandian continent-continent collision (**425–400 Ma**) which led to at least a doubling of the crustal thickness by "subduction" of the Western Gneiss Complex beneath the Laurentian Plate (Greenland), and formation of eclogites within the "subducted rocks". During this orogenic phase, the bedding-parallel $\mathbf{S_1}$-foliation in the Høydalsfjorden Complex was established, with an $\mathbf{M_1}$-metamorphism corresponding to a prograde lower greenschist facies (upper chlorite-grade). The $\mathbf{S_1}$-foliation is locally related to shear movements, and this shear fabric has been related to the Scandian eastward thrust-emplacement of the rocks. The $\mathbf{S_1}$-fabric was folded into $\mathbf{F_2}$-folds having W-E to WNW-ESE trends and shallow to steep plunges, with the folds showing either "curved" fold closures or kink/chevron forms. The axial planes strike WNW-ESE and have mainly steep dips. A penetrative axial planar $\mathbf{S_2}$-cleavage has not been developed, but $\mathbf{S_2}$-fabrics are represented by kink bands and locally by crenulation cleavage. The $\mathbf{S_2}$-fabric displays only vague mineral recrystallisation, but the mineral assemblage appear to be the same as for the $\mathbf{S_1}$-fabric, corresponding to lower greenschist facies (upper chlorite grade). The $\mathbf{F_2}$-folds are cut by the sub-Devonian unconformity, and the $\mathbf{D_1}$- and $\mathbf{D_2}$-deformations are thus pre-Devonian. No marker lithologies exist in the Høydalsfjorden Complex, and thus the internal stratigraphy, as well as the overall structural geometry of the rocks cannot be unravelled.

It has been discussed whether Scandian effects may also be found in the rocks of the <u>Eikefjord Group (EG)</u>. In these respects, the remnants of the strongly retrograded ***amphibolite facies*** mineralogy in the $\mathbf{S_2}$-mylonites, would be a candidate for Scandian tectonometamorphic features. The amphibolite facies mineralogy may be a result of four alternative processes: **(1)** *Amphibolite facies due to "Jotun Nappe-type" retrogression*. This alternative implies a general shallow- to middle-crustal retrogression of the original Precambrian granulite facies, similarely to what has been reported from the Jotun Nappe and the Dalsfjord Nappe. However, since the Eikefjord Group, during the Scandian, appears to have experienced a higher grade of metamorphism than the Jotun and Dalsfjord nappes, the “Jotun Nappe-type” retrogression is considered <u>less likely</u> as an explanation for the amphibolite facies. **(2)** *Amphibolite facies due to “down transport” related to subduction of the WGR.* In this alternative, the Eikefjord /Lykkjebø rocks are assumed to have been transported to lower-/sub-crustal conditions during subduction of the WGR, producing a prograde mineral assemblage. Johnston et al. (2007b) made similar suggestions. This scenario would give an amphibolite facies metamorphism of <u>Scandian age</u>. **(3)** *Amphibolite facies due to movements along, and exhumation of, the NSDZ.* Here it is assumed that the amphibolite facies metamorphism was a result of exhumation of the WGR along the NSDZ, implying exhumation also of the NSDZ itself. In this scenario, the amphibolite facies could itself be just a stage in the retrogradation of a higher grade rock. However, Johnston et al. (2007b) claimed that the upper amphibolite facies represented the *peak* of the prograde metamorphism. If so, the rocks were never deeper than the upper amphibolite facies level. If the amphibolite facies minerals formed during exhumation of, and shear along, the NSDZ, the amphibolite metamorphism may be classified as *post-Scandian*. **(4)** *Amphibolite facies due to* *heating from exhuming hot eclogites below the NSDZ.* In this alternative, the prograde amphibolite facies metamorphism is seen as a result of heating from the hot eclogites/WGR rising below the NSDZ. The NSDZ is a Mode-II feature, associated with post-Scandian extension related to plate divergence between Laurentia and Baltica. Hense, a n amphibolite facies caused by heat from the rising of the WGR/eclogites below the zone would be a post-Scandian feature.

In summary, it appears that only model (2) "*Amphibolite facies due to "down transport" related to subduction of the WGR*", will give a Scandian age for the amphibolite facies metamorphism.

Post-Caledonian, Devonian, extensional geological history

The Devonian, extension-related geological history in the study area is well documented by the formation and deformation (folding) of the Håsteinen Devonian Massif (HDM). Also the rocks of the Eikefjord Group (EG) have been severely effected by these Devonian tectonic processes, which led to penetrativ top-to-the-west mylonitisation of the rocks, as well as folding. Regarding the Høydalsfjorden Complex (HC), zones of top-to-the-west mylonites occur only locally, whereas the effects of folding is more obscure. Hence, two tectonic styles have evolved in the area during the Devonian geological history: first the onset of top-to-the-west extensional movements, then the onset of folding.

Top-to-the-west Devonian extensional movements in general. At the peak of the Caledonian Orogeny in the Early Devonian, the contractional movements of the orogeny ceased, and were immediately replaced by crustal-scale extensional movements. The first stage of extension, occurred by westward movement of the entire orgenic wedge on the basal décollement zone (**Mode-I** extension of Fossen 1992, 2000; Milnes et al. 1997; Fossen & Dunlap 1998). Subsequently, in western Norway, the continued extension led to the development of a westward dipping crustal-scale detachment zone that was separating an "Upper Plate" in the west from a "Lower Plate" in the east. The detachment zone, called the Nordfjord–Sogn Detachment Zone (NSDZ), cut through the nappe stack, and may be viewed as a fundamental crustal-scale decoupling of the orogenic belt. The movements on this zone (**Mode-II** extension of Fossen 1992, 2000; Milnes et al. 1997; Fossen & Dunlap 1998) appear to have controlled the formation of the Devonian basins in western Norway. In the literature, several models have been suggested to explain the uplift of the eclogitic Lower Plate and the onset of extension. Andersen & Jamtveit (1990) suggested a model where these processes were caused by orogenic collaps resulting from removal of the "thermal boundary layer" (orogenic root) by its descent into the lithosphere. Later workers, however, have suggested the exactly opposite movement of the root, i.e. that the uplift was in fact a result of the exhumation of the *preserved* orogenic root due to buoyancy (Milnes et al. 1997). The extension has also been explained by large-scale sinistral wrench tectonics (e.g. Osmundsen & Andersen 2001). Fossen & Dunlap (1998) dated the **Mode-I** westward movement of the entire orogenic wedge to **402–394 Ma**, and concluded that the **Mode-I** and **Mode-II** extension was a result of reversal of the Baltica–Laurentia plate movements from convergent to divergent (Fossen 2000).

In the present thesis, a mismatch has been revealed to exist between the $^{40}Ar/^{39}Ar$ muscovite ages of the NSDZ in the Eikefjord–Gloppen area, and the status of the NSDZ as a Mode-II zone. The $^{40}Ar/^{39}Ar$ muscovite ages in the Eikefjord–Gloppen area fall in the range **404–398 Ma**, which is contemporaneous with the movements at **402–394 Ma** on the Mode-I zone of Fossen & Dunlap (1998). But, in addition, the NSDZ shows extensive movements at amphibolite facies conditions, indicating that the movements on the NSDZ started even earlier than the time interval **404–398 Ma**. Recently, Johnston et al. (2007b) performed P/T analyses and radiometric dating in the Eikefjord area, and interpreted the NSDZ to have moved in the time intrval **410–400 Ma**. Viewed together, these ages may seem to indicate that the Mode-II sone developed *before* the Mode-I zone, i.e. oppsosite to the conventional order of appearance. However, as the establishment of the "Mode-I first" and "Mode-II next" is based on independent field work (Fossen 1992, 2000; Fossen & Dunlap 1998; Milnes et al. 1997), the conventional order of appearance has to be maintained. The causes of the mismatch is not clear.

Effects of Devonian extensional movements, on the Høydalsfjorden Complex: (*top-to-the-west movements predating deposition of the HDM)*. The post-Scandian extensional movements led to formation of top-to-the-west $\mathbf{S_3}$-mylonites in the Høydalsfjorden Complex (HC). In the study area, such mylonites are present for example at the north margin of the Høydalsfjorden rocks, i.e. near the Sunnar Fault, and at Svardal near the southern margin of the study area, i.e. near the Standalen Fault. In the HC rocks, a very limited number of folds with NE-SW trending fold axes have been recorded. The trend of these folds stands with a high angle to the abundant WNW-ESE trending HC $\mathbf{F_2}$-folds that are cut by the unconformity below the Håsteinen deposits. The NE-SW trending folds might be $\mathbf{F_3}$-folds related to the $\mathbf{D_3}$-top-to-the-west displacement, although the limited data makes it uncertain whether the folds are really $\mathbf{F_3}$-folds. The top-to-the-west $\mathbf{S_3}$-mylonites are cut by the unconformity present below the Devonian Håsteinen sediments. This suggests that the $\mathbf{S_3}$-mylonites formed at an early stage during the Devonian extension, *prior* to the formation of the Devonian basin. Since the formation of the Håsteinen basin was probably related to the movements on the **Mode-II**/NSDZ/EG, the HC $\mathbf{S_3}$-mylonites appear to have *predated* the major part of the movements along the Eikefjord Group mylonites. Hence, it appears that the HC $\mathbf{S_3}$-mylonites are not related to the Eikefjord Group mylonites of the **Mode-II**/NSDZ. This is consistent with the lower degree of metamorphism in the HC $\mathbf{S_3}$-mylonites, as evidenced by the presence of brittle quartz, suggeting temperatures *below* greenschist facies. Instead, the HC $\mathbf{S_3}$-mylonites may be related to the foregoing **Mode-I** extensional phase.

Although the Høydalsfjorden unit does not appear to have been affected by **Mode-II**/NSDZ/EG-type of shear at the presently exposed structural level, such shear may be present at greater depths. If the Sunnar Fault – which separates the Høydalsfjorden Complex to the south, from the Eikefjord Group to the north – has been the locus of kilometre-scale dip-slip displacements with downward movement of the south block (Høydalsfjorden rocks), then it is possible that the Høydalsfjorden block might contain Eikefjord-type mylonites at a deeper structural level, that is now down-faulted by the Sunnar Fault.

Effecs of Devonian extensional movements, on the Eikefjord Group. In the study area, the Nordfjord–Sogn Detachment Zone (NSDZ) is developed in the rocks of the Eikefjord Group. In this group, the extensional $\mathbf{D_2}$-movements produced the strong and dominating penetrative mylonitic $\mathbf{S_2}$-fabric which generally obliterates all older structures and is retrogressive from amphibolite facies (mentioned above) down to middle or lower greenschist facies. The amphibolite facies might originally have formed as a result of partly "down-transport" of the EG rocks during subduction of the WGR. A contribution to the last stages of such a prograde metamorphism might have come from heat transferred from eclogitic rocks that were present in the Lower Plate just below the NSDZ. The metamorphic retrogression of the EG rocks from amphibolite to greenschist facies reflects the uplift of the rocks during the extensional movements on the NSDZ. Shear sense indicators in the mylonite show that the mylonite-producing extensional movement has been of top-to-the-west type. The related $\mathbf{L_2}$-stretching lineations have a WNW-ESE trend and shallow plunges.

Formation of the Håsteinen Devonian basin. Deposition of sediments. The basin-formation and the deposition of Devonian Old Red Sandstone (ORS) sediments were generally controlled by the extensional movements that led to west- and downward movement of the Upper Plate and gradual uplift of the Lower Plate. The sediments were accumulated on top of the Upper Plate, i.e. on the Høydalsfjorden Complex, covering the latter complex and its lower greenschist facies mineralogy and the accompanying $\mathbf{D_1}$- and $\mathbf{D_2}$-structures that were related to Scandian top-to-the-*east* orogenic transport, as well as its sub-greenschist facies $\mathbf{D_3}$-mylonites related to top-to-the-*west* extension. The Devonian sediments themselves consist of conglomerates and very minor sandstone units, deposited in a continental setting. The conglomerates were deposited by mass flow processes in a proximal alluvial fan environment, whereas the sandstones were deposited by fluvial processes in a more central or distal part of the fan environment. The material was deposited with a primary unconformable contact on the Upper Plate rocks, and this unconformity is exposed at numerous places around the Håsteinen Devonian Massif (HDM). The unconformity shows that the HDM is situated *in situ* with respect to its immediate substrate.

In the E-W profile going through the centre of the present-day Håsteinen deposits, the beds are standing with an eastward dip of **53°** against the subjacent Høydalsfjorden rocks, and the unconformable contact between the two units forms a subhorisontal surface. This means that the Devonian beds, which must originally have been deposited *sub-horizontally*, were rotated to a *steep eastward dip* during the extensional movements. Likewise, due to the primary unconformity, the basin floor, defined by the Upper Plate (HC), must have rotated accordingly, from an original steep *westward* dip at the time of ORS deposition, to the present *subhorisontal* orientation. The *present-day* **53°** angle between the steeply dipping bedding and the overall subhorisontal contact towards the substrate, represents an extraordinary geometry. This was also the case with the high angle that must have existed at the *time of sedimentation*, when the sub-horizontal Devonian bedding was deposited against a steeply westward-dipping palaeo-slope.

The extraordinary geometry, defined by the large bedding/contact angle that existed in the Håsteinen basin at the time of deposition, cannot be explained by the hitherto published models for the formation of the west Norwegian Devonian massifs. The hitherto published models suggest that the sediments were deposited directly towards the fault plane of a constantly active (listric?) growth fault, a situation implying that the bedding/substrate contact would be a tectonic fault, and not a primary unconformity as in the HDM. A new model is therefore required to explain the data in the HDM.

Such a new model – termed the **ramp basin model** – has been presented in the present work. The unique feature of the model is that the deposition is believed to have occurred in a depression that formed on top of the Upper Plate as a result of a westward-dipping frontal ramp in the subjacent detachment zone. This is in contrast to previous models in the literature, which suggest that the deposition occurred in a basin of half-graben type. In the ramp basin model setting, such a half-graben basin would form at a considerable distance *east* of the “ramp basin”, in a position above the listric detachment zone present just west of the cut-off line of the detachment zone. The Hornelen Devonian deposits appear to have formed in a half-graben basin of the latter type.

The ramp basin model has been tested by **numerical modelling**, in an attemt to reproduce the crustal-scale extensional development of the Håsteinen and adjacent area. The modelling has been carried out by means of the balancing/restoration/forward modelling program “**2D-Move**”. In the thesis, the technique of “**foreward modelling**” has been applied. Several E-W trending vertical cross-sections have been produced to show a step-by-step development of the crustal-scale extension and basin formation.

The model has several characteristic features. The ramp in the detachment divides the Upper Plate into a thin eastern part and a thick western part, that lies on the detachment zone (the Nordfjord–Sogn Detachment Zone). From west to east, the detachment zone itself can be divided into four segments: lower flat, ramp, upper flat, and eastern fault – of which the eastern fault reaches the daylight cut-off point. In the Sunnfjord Region, the ramp, as well as the related ramp depression, must have been located *east* of the Håsteinen sediments, yet at a position lying at a considerable distance *west* of the original eastern termination of the Upper Plate (cut-off line), although the original position of the ramp is not known. During the modelled extensional movements, the thin part of the Upper Plate, positioned on the upper flat of the detachment zone, is continuously moving down the ramp slope. Above the ramp, a ramp depression (ramp syncline) is developed on the surface of the Upper Plate, forming a sedimentary basin. Subhorisontal Devonian beds are deposited in the depression, and the angle between the bedding and the westward-dipping sub-Devonian contact will be substantial. The contact itself will be a primary depositional unconformity. During the movements of the sub-horizontal bedding down the ramp-slope, the bedding will, as a geometrical necessity, gradually rotate to aquire a *steep eastward-dip*. At an advanced stage, a part of the <u>thin</u> Upper Plate will be lying (sub-)horisiontally on the lower flat, and Devonian bedding will be standing on top with a steep eastward dip. A few major features of the cross-sections will be briefly summarised in the following.

Dip of detachment. Hanging wall shear angle. In the modelling, two series of cross-sections have been produced: a first series where the dip-angles of the “lower flat–ramp–upper flat–eastern fault” is **0–35–0–25°**, and where vertical isostacy has *not* been applied – and a second series where the corresponding angles are **5–40–0–25°**, and where isostatic movements *have* been calculated. In both series, the shear angle in the hanging wall block has been set to **90°**. This angle controls the accomodation of the hanging wall block onto the shear surface during the movements of the block.

Basin margin uplift. In the latter series, it is seen that the Hornelen-type half-grabens experience basin-margin uplift, whereas the ramp-basin does not. The absence of basin margin uplift at the ramp basin is due to the steeper angle of the ramp-segment of the detachment, and an increased sediment load there.

Bedding dip of 35° instead of 53°. The ramp has been modelled with a dip-angle of **35°**, instead of the dip of **53°** which is present along the axial trace of the Osstrupen syncline. The dip of **35°** has been obtained from a comparison with the Hornelen Devonian massif, where the beds are generally dipping **25 °E** throughout, but where the folding, that formed the Grøndalen syncline, led to a steepening of the bedding to **38 °E** along the axial trace of the Grøndalen syncline. This is a steepening of **52 %**. A similar steepening has been assumed for the folding of the Håsteinen deposits, implying that the Håsteinen beds were dipping **35 °E** prior to folding.

Relationship between general isostatic uplift and dip of the detachment. In the second series where the isostacy has been applied, two segments of the detachment zone becomes uplifted. One of the segments is located underneath the Hornelen-type half-grabens in the east. This uplift is part of the basin margin uplift, but also gives the detachment zone a westward dip, facilitating westward movement of the Upper Plate. The other segment is defined by the entire lower flat (i.e. west of the ramp), where the segment-part that lies just below the Håsteinen deposits is lifted and rotated to a subhorizontal orientation, whereas the segment-part that lies west of this is uplifted while maintaining its dip of **5°**.

In an evaluation of the numerical modelling, it appears that the modelling succeeds in reproducing the geolgical situation in the Håsteinen area.

In the Sunnfjord Region, the Devonian extensional process and the presence of a ramp basin eventually managed to produce the bedding-normal, cumulative stratigraphic thickness of **5.8 km** for the HDM, as calculated along the E-W trending axial trace of the Osstrupen syncline. Also the Håsteinen sediments that are located east of the study area, as well as the sediments that were probably once located west of the present-day massif, can be explained by the model. Hence, the ramp basin model explains all the important data concerning the formation of the HDM.

Devonian folding in general: Folding of Devonian age has affected all three units in the study area. In all the units, the fold axes are trending WNW-ESE, indicating a roughly NNE-SSW directed contration. In the literature, the folding of the Devonian deposits and subjacent units in Western Norway have been interpreted as a result of a regional contractional event, affecting the entire area between Bergen and Trøndelag. In the present thesis, however, it is suggested that the folding of the Håsteinen basin, and its substrate, has mainly been a result of intrabasinal or local processes related to the extension, indicating that regional N-S contraction may not be necessary to explain the folding.

Devonian folding of the Håsteinen Devonian Massif (HDM). In the HDM, the Devonian folding led to formation of the Osstrupen $\mathbf{F_1}$-syncline. This syncline developed into an upright, plane fold with an axial trace oriented WNW-ESE (**295°–115°**) and with the average fold axis having a ESE trend and a moderate to steep ESE-ward plunge (**115/53**). The orientation of the fold axes is remarkably constant along the axial trace, deviating less than **+/- 10°** from the average. The limbs are straight, with a mean strike/dip orientation **070/62 SE** for the northern limb and **161/62 NE** for the southern limb, and an interlimb angle of **103°**. A hinge zone is only present in the northern fold-half, i.e. north of the axial surface. In the Gravaneset sandstone unit, which is located in the westernmost parts of the HDM, a set of first order parasitic $\mathbf{F_1}$-folds with related axial planar $\mathbf{S_1}$-cleavage have been developed. The lack of systematically distorted clasts in the HDM may suggest that some of the folding occurred as soft sediment deformation. However, the presence of the axial planar cleavage at

Gravaneset shows that the major part of the deformation occurred during semiductile/ductile (plastic) conditions. In the thesis, three models have been presented to explain the folding, and these are summerised in the folowing.

Fold model 1: folding due to transpression along converging lateral ramps ("Lateral ramp" model). In this model, the NSDZ forms a trough between two roughly E-W trending lateral ramps, that are not parallel. The ramps gradually converge towards the west, and the Upper Plate (Høydalsfjorden Complex) forms a depression that mirrors this trough. The westward movement of the Upper Plate, with Devonian sediments on top, may cause the two units to fold between the converging lateral ramps. This might happen on the condition that the rock volume, which is initially located between the two ramps, essentially maintains its position between the ramps. However, as the validity of this condition may be questioned, the model is problematic.

Fold model 2: folding due to ridge-shaped frontal ramps ("Ridge-ramp" model). In this model, the folding occurs just above the westward-dipping ramp in the NSDZ. During its westward transport, the Upper Plate moves down the ramp. The ramp then controls the formation of the ramp basin, into which the Devonian beds are deposited horisontally on top of the Upper Plate. During the movement of the beds and the Upper Plate down the ramp, the beds aquire their eastward dip. In the present model, the unique feature is that the ramp is not planar, but formed as a ridge, with the "crest-line" of the ridge plunging in the down-dip direction. As a geometrical necessety, the downward movements of the beds will lead to formation of a fold structure. The model succeeds in explaining the formation of the Osstrupen syncline.

Fold Model 3: folding due to narrowing basin at depth ("Narrowing basin depth" model). In this model, the folding again occurs above the westward-dipping ramp in the NSDZ, as in Fold model 2 above. However, in contrast to Fold model 2, the ramp is now planar. The beds deposited on the Upper Plate is once again transported down the westward-dipping ramp, whereby they achieve their eastward dip. However, the folding is now controlled by the dip of the northern and southern margins. These margins are dipping southward and northward, respectively, and this means that the basin is much wider at the daylight surface than at the bottom. During the downward movement of the beds, they experience a basin-width that is gradually reduced, and the beds will have to fold passively into a syncline. The model therefore provides a simple explanation for the formation of the Osstrupen syncline. When the fold has formed, it may be influenced by the movements along lateral ramps.

In an evaluation of the three models, it appears to be logical that the basin-width will be reduced at depth, making Model 3 a likely explanation. It is possible that a ridge-formed ramp has contributed to the folding, but a very pronounced ridge-ramp morphology combined with reduced basin-width at depth could easily lead to "double"-folding, i.e. one fold from each process, and this is not observed in the study area. Hence, it seems most likely that the major part of the folding is a result of the reduction of basin width towards depth – although the fold may have been affected by subsequent movement along lateral ramps.

The Devonian folding eventually produced the present-day orientation of the unconformable contact below the HDM. In the western parts of the massif, where the contact along the northern and southern margins clearly have massifward dips, the unconformity underneath the HDM is interpreted as a trough (i.e. synclinal form). In the eastern areas, the presence of substrate inliers within the HDM implies that the contact underneath the Devonian sediments is more irrregular. Here, the massifwide envelope surface, defined by the contact, has been interpreted as subhorisontal.

<u>*Devonian folding in the Høydalsfjorden Complex (HC)*</u>. In the Høydalsfjorden Complex, the effects of Devonian N-S folding are more uncertain. The major HC $\mathbf{S_1}$-foliations – and the HC $\mathbf{F_2}$-fold structures which trend roughly parallel to the $\mathbf{F_1}$-fold axis of the HDM Osstrupen syncline – are cut by the unconformity below the Devonian deposits. Also the top-to-the-west $\mathbf{S_3}$-mylonites are cut by the sub-Devonian unconformity. Since these structures formed prior to deposition of the Devonian sediments, they are not candidates for Devonian folding. A small number of fold axes are trending NE-SW, i.e at a high angle to the $\mathbf{F_2}$-folds. These NE-SW trending folds are possibly $\mathbf{F_3}$-folds, although their relationship to the $\mathbf{S_3}$-mylonites and the unconformity, have not been observed.

However, candidates for Devonian folding are possibly present locally, in the orientation of the $\mathbf{S_3}$-mylonite fabrics. These fabrics give the impression of having variable dips towards the SSW or NNE, possibly suggesting that the fabrics are folded around fold axes trending WNW-ESE. These possible folds have been termed $\mathbf{F_4}$-folds, and may have formed during the N-S contraction that also folded the Håsteinen deposits.

Generally, it is difficult to reveal the possible effects, that the N-S contraction of the HDM had on the Høydalsfjorden complex. This is due to the fact that the trend of the HC-$\mathbf{F_2}$-folds that formed prior to Håsteinen deposition, is roughly parallel to the trend of the later HDM-$\mathbf{F_1}$-Osstrupen syncline. It is possible that the HC-$\mathbf{F_2}$-folds have been somewhat tightened during the Håsteinen contraction. In additon, it cannot be excluded that the Høydalsfjorden Complex actually contains folds related to the Håsteinen N-S contraction, but that the folds have been recorded as HC-$\mathbf{F_2}$-folds due to the geometrical concordance with the latter folds. One could, for example, speculate as to whether some *kink* folds in the Høydalsfjorden Complex, which have been designated $\mathbf{F_2}$, were formed during the N-S contraction. If the Håsteinen sediments were folded according to “Fold model 3” above (“Narrowing basin depth” model), then the HC-unit could have been folded by passive bending that would give little deformation in the unit.

The only place where it can be clearly seen that the Devonian N-S contraction has affected the Høydalsfjorden rocks on the mesoscopic scale, is at Gravaneset in the far west. Here, the unconformity below the Gravaneset Sandstone Unit is parallel to the sandstone beds. The folding of the sandstone beds has thus led to a corresponding folding of the unconformity. However, the Høydalsfjorden rocks below the unconformity do not appear to have any particular structures, separable from the pre-unconformity $\mathbf{F_2}$-folds, etc., that can be related to the folding of the unconformity.

<u>*Devonian folding in the Eikefjord Group (EG).*</u> In the Eikefjord Group, the Devonian contraction folded the $\mathbf{S_2}$/$\mathbf{L_2}$-mylonite fabric into $\mathbf{F_3}$-folds with fold axes having a WNW-ESE oriented trend and shallow plunge, making the $\mathbf{L_2}$-stretching lineations and the $\mathbf{F_3}$-fold axes parallel. An $\mathbf{S_3}$-axial planar fabric was not developed. Furthermore, the trend of the $\mathbf{F_3}$-folds in the Eikefjord Group is parallel to the trend of the Osstrupen $\mathbf{F_1}$-syncline in the HDM. The envelope surface of the ($\mathbf{F_3}$-)folded $\mathbf{S_2}$-mylonite fabric has a semihorisontal to shallow southward dip.

The Devonian N-S contraction produced the **F_3**-folding of the **S_2**-mylonites of the Eikefjord Group, since the **F_3**-folds are folding the extensional **S_2**-mylonites. The parallel orientation of **L_2**-stretching lineation and **F_3**-fold axes in the Eikefjord Group might indicate a genetic relationship, although this is still somewhat uncertain. Since the **F_3**-folding in the Eikefjord Group appears to be more intence than the **F_1**-folding giving the Osstrupen syncline, the **F_3**-folding could be a result of shear-related contraction within the detachment zone.

Post-Devonian geological history

A large number of faults and joints are developed in the HDM. The amount of displacement has been tested for the most prominent faults crossing the massif, and the movements are found to be negligible. A minor fault at the Devonian-substrate contact below the Gravaneset sandstone unit has been dated by palaeo-magnetism to a Triassic/Early Jurassic age (Torsvik et al. 1987).

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