SOME SEDIMENTOLOGICAL AND STRUCTURAL STUDIES OF THE OLD RED SANDSTONE HITRA GROUP, HITRA, SØR-TRØNDELAG

VOLUME I

TEXT

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ABSTRACT

The Old Red Sandstone (O.R.S.) area on Hitra contains an about 1350 m thick, continental succession of conglomerates, sandstones and mudstones, here referred to as the Hitra Group. It has further been divided into 3 formations, from bottom to top the Aune Formation, The Vollan Formation and the Balsnes Formation. These have again been divided into Members A-L.

The group commences with a basal breccia overlain by floodplain mudstones and sandstones. This in turn is overlain by braided river conglomeratic sandstones followed by a cyclic sequence of mudstones and sandstones interpreted as representing floodplain/distal alluvial fan and low sinuosity river deposits. The remaining part of the sequence is an alternation of alluvial fan conglomerates and sandstones, temporary lake mudstones, braided and meandering river sandstones, mudstones and conglomerates and finally a coarse alluvial fan conglomerate.

The sediments were laid down in an area which periodically may have linked up with other O.R.S. areas in the region. Deposition of the Aune and Vollan Formations took place towards the NE, E and SE, while the Balsnes Formation was deposited towards the SE, S and SW. Repeated uplift and rejuvenation of source areas, possibly along marginal faults, resulted in progradation of alluvial fans into more distal parts of the depositional basin, normally characterized by mudstones and sandstones.

Tectonic deformation folded the succession into two, major, ENE-WSW trending synclines with higher order parasitic folds on the limbs. There is a general noncylindrical style of folding with transitions into aberrant folds. This first phase of deformation, which also includes locally strong cleavage development and thrusting, is regarded as of lowermost Upper Devonian age.

Later deformation resulted in N-S trending kink bands and monoclines, E-W trending crenulation cleavage and locally strong faulting. Fault directions are dominantly NE-SW and NW-SE. These structures can probably be related to Permo-Jurassic and later crustal movements. The succession was subjected to low-grade metamorphism during burial and deformation.

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SYMBOLS AND ABBREVIATIONS

LOG SYMBOLS

 Parallel Lamination and Horizontal Bedding

 Trough Cross-Bedding

 Tabular Cross-Bedding

 Lenticular Lamination

 Ripple Lamination

 Climbing Ripple Lamination

 Wave Ripple Lamination

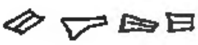
 Bioturbation

 Flame Structures

 Ball and Pillow structures

 Flaser Bedding

 Calcareous Concretions

 Intraformational Mudclast Conglomerate

 Gradual Transition between Beds

 Erosive Scour at the base of a bed

 Current Direction Indicator

GRAIN SIZE ABBREVIATIONS ON LOGS

cl	Clay
s	Silt
vf	Very Fine Grained Sandstone
f	Fine Grained Sandstone
m	Medium Grained Sandstone
c	Coarse Grained Sandstone
vc	Very Coarse Grained Sandstone
gr	Conglomerate

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LIST OF 1:5000 "ØKONOMISK KARTVERK" MAPS USED FOR MAPPING

NAME	NUMBER
Hamnalangvatnet	BR 128-5-2
Balsnesaunet	BR 128-5-3
Balsnes	BR 128-5-4
Hamnadalen	BS 128-5-1
Akset	BS 128-5-2
Langneset	BS 128-5-3
Heilfjellet	BS 129-5-4
Terningen	BT 128-5-1
Strand	BT 129-5-2
Fleinskallen	BT 129-5-3
Sandstad	BT 129-5-4
Vedøya	BU 129-5-1
Borøysund	BU 129-5-3

C H A P T E R 1

INTRODUCTION

1 INTRODUCTION

1.1 PRACTICAL INTRODUCTION AND PURPOSE

The "Devonian" rocks of Western Norway, Røragen and Trøndelag are regarded as post-Scandian or post-orogenic molasse deposits, i.e. equivalent to Old Red Sandstone (O.R.S.). Thus the author will refer to these Siluro-Devonian molassic facies as "Devonian" for purposes of simplicity. The age problem will be further discussed in Chapters 4.7 and 7.1.

Earlier two studies have been carried out on the "Devonian" on Hitra, (Peacock, 1965; Siedlecka and Siedlecki, 1972). In addition Siedlecka (1975) examined the calcareous concretions in the sediments. Nobody put much weight on describing the structural geology of the area.Therefore one aim of this project has been to remap the whole "Devonian" area on Hitra.

The structurs were examined in detail in order to figure out the structural pattern and sequence of deformation. The area was mapped on a scale of 1:5000 to get a high degree of accuracy. All the data were transfered onto 1:15000 maps constructed from air photographs (Plates 1A and 1B). Structural data were transfered onto overlays (Plates 2A and 2B).

The 1:5000 maps belong to the series "Økonomisk Kartverk". All grid references refer to these maps. Grid references are given by two numbers. The N-S coordinates are given by the first numbers, while the second numbers give the E-W coordinates. Finally, all names referred to in the text are to be found on the 1:15000 maps (Plates 1A and 1B).

The second aim of this project was to make logs through the sedimentary sequence and try to interpret facies and environments of deposition. This had been done earlier by the above mentioned workers, but not in great detail.

This thesis is based on field work done during the summers of 1984 and 1985. A total of three and a half months were spent in the field.

A total of about 100 thin sections were examined during the work.

Financial support has been provided by the University of Bergen and Elf Aquitaine Norge A/S through the "Devonian Project".

1.2 EARLIER WORK ON THE OLD RED SANDSTONE SUCCESSION ON HITRA

The first person to make a good description of the "Devonian" sediments on Hitra was Reusch (1914). Earlier work by Schetelig (1913) was mainly concentrated on the age relationships between the sediments and the underlying diorite . He concluded that the diorite was the younger, which Reusch was able to show was wrong. Later Vogt did a lot of work in the area during the period 1923-1929 (Vogt, 1928; Vogt, 1929).

Little was then done until 1965 when Peacock made a detailed sedimentological study (Peacock, 1965), and in 1972 Siedlecka and Siedlecki published a paper on sedimentary and structural geology on the Downtonian on Hitra (Siedlecka and Siedlecki, 1972). They concluded that the sediments represent deposition on alluvial fans, in braided and meandering rivers, lakes and overbank areas, and that the succession is folded in a large, open, ENE-WSW trending syncline. Siedlecka (1975) also attempted to correlate the "Devonian" areas in Outer Trøndelag with another.

Siedlecka (1977) published another paper on calcareous concretions in the area. A study on tectonic deformation was carried out by Roberts (1981), who also made a compilation of his and earlier workers' conclusions on the "Devonian" (Roberts, 1983). The most recent paper on the O.R.S. on Hitra is that by Steel, Siedlecka and Roberts (1985). It summarizes the investigations up to the present day.

C H A P T E R 2

GENERAL INTRODUCTION ON THE "DEVONIAN"

2 GENERAL INTRODUCTION ON THE DEVONIAN

2.1 GENERAL

To achieve a broader understanding of the structural and sedimentological development in the field area, it was considered necessary to make a short overview of other sediments of the same age in the North Atlantic Region. Prior to the opening of the Atlantic Ocean in Early Tertiary times, Norway was situated close to the North American craton, with Greenland as its neighbour to the west (Fig. 2.1). Therefore a brief summary of the Devonian in that area is considered important. Summaries will also include the Devonian sequences of Scotland, Spitsbergen and other areas of Norwegian "Devonian" rocks.

2.2 EASTERN GREENLAND

Devonian sediments in Easteren Greenland were deposited during Middle and Upper Devonian times, and occur over a relatively large area, about 150 km long and 60 km broad. Maximum thickness is over 7000 m in the middle of the area. In addition there are isolated outcrops both to the north and to the south, for example on Canning Land (Haller, 1971).

The main Devonian outcrop is cut by a prominent fault line with downthrow to the E. East of this line the Devonian is covered by younger rocks. The western boundary defines the original trough margin for the Middle Devonian, while the Upper Devonian sediments have extended further westwards. The Devonian sediments rest unconformably on strongly folded Cambro-Ordovician sediments and rocks belonging to the Eleanore Bay Group. These rocks were deeply eroded before deposition of Devonian rocks.

During Devonian times Eastern Greenland was an intramontane basinal area with deposition of late molasse sediments in a fairly warm and arid climate. Tectonism led to steadily changing depositional environments, both vertically and horizontally. As a result , the composition of clastics changed as boundaries of supply areas shifted (Haller, 1971).

The stratigraphic succession is divided into cycles separated by disconformities (Bütler, 1959; Haller, 1971). The cycles are separated by disconformities indicating tectonic events. Every new cycle then starts with a basal conglomerate. In general the Middle Devonian is a basal conglomerate with breccia at the bottom . In the Upper Devonian there is a greater variation in sedimentary environments, e.g. deposition by rivers, in lakes, lacustrine deltas, on floodplains and alluvial cones along the margins of the basin and in connection with tectonically induced rises. No sign of marine influence has been detected. The sediments are generally poorly sorted and vary in colour from deep red through green to grey with all sorts of alternations. Sandstones were deposited on wide floodplains, conglomerates in deltas and around the margins of smaller basins, while breccias were deposited from mountainous terrains. The coarser sediments are often cemented by carbonates. Occasionally marls and calcareous beds were deposited in quiet basins.

Volcanic activity was widespread during deposition of the Devonian in Eastern Greenland. The volcanic centres are scattered, and were active at different stages. Both mafic and silicic lavas are common, the former lavas being calc-alkaline, the latter calc-alkaline and alkaline (Bütler, 1959). Volcanicity followed periods of late orogenic uplift and postdates the main orogenic episodes. It is believed to be associated with late orogenic "spasms". Volcanic vents, dikes and flows are common in the molasse deposits. In addition to volcanicity, emplacement of late orogenic granitic plutons was widespread. They are classified as syn-tectonic intrusions and post-tectonic batholiths and stocks (Haller, 1971). The latter category is closely tied to the late orogenic "spasms". Devonian dykes and pegmatite swarms are found cross-cutting Devonian sediments.

Two fracture systems can be discerned. The oldest one is a set of tensional fractures striking NNW-SSE. These originated just after the main period of mountain building. Crustal stretching of up to 30 % has been calculated (Haller, 1971). An "en echelon" fault pattern developed, and

Devonian sediments were deposited in grabens. Later a set of high angle normal faults developed. They strike N to NNE, but there are E-W faults as well. There is a consistent eastward downthrow along longitudinal faults, and movements of several thousand meters has been recorded.

Four phases of intra-Devonian movements are recorded, two Middle Devonian and two Upper Devonian (Bütler, 1959). Both folds and thrusts developed. One set runs NNW to N and involves thrusting of basement slices through the Devonian succession. This phase is of Middle Devonian age. In the Upper Devonian deformation took place again with folds and thrusts running NE-SW. The earlier formed horst and graben structures were then strongly deformed. Many of the longitudinal NNE-SSW faults were transformed into reverse faults and overthrusts. Even collapse slides formed (Haller, 1971; Escher and Stuart Watt, 1976).

2.3 SPITSBERGEN

The Devonian sediments on Spitsbergen are found in the Central Province (Harland and Wright, 1979). The largest outcrop area is in the north, where the outcrops measure about 150 km x 75 km and the succession is about 8 km thick. The sequence rests unconformably on similar looking Silurian sediments, which again rest on Caledonian metamorphics. To the W and E the Devonian is bounded by Upper Devonian sinistral strike-slip faults (Friend and Moody-Stuart, 1972), to the S it is buried below younger rocks. Another outcrop of Devonian rocks is found further south at Hornsund.

The succession has been divided into the Red Bay Group and the Andree Land Group. The Marietoppen Group near Hornsund is correlated with the latter (Harland and Wright, 1979). These are subdivided according to faunal differences. The succession ranges in age from Late Downtonian to Early Frasnian (Harland and Wright, 1979). Similar faunas have been encountered in Eastern Greenland, North-Eastern America and Britain. Some Upper Devonian sediments are preserved, which were deposited after the Svalbardian Orogeny. They continue into the Carboniferous without breaks.

The Devonian grey-red continental and partly marine sediments on northern

Spitsbergen have by a number of authors been interpreted as representing deposits laid down during an extensional tectonic phase within an extensive graben. They accumulated in a wide variety of environments, for example on alluvial fans, in meandering and braided rivers, on playas, in lakes and in deltaic areas. All types of sediments usually found in such environments are common, e.g. conglomerates, sandstones, siltstones, mudstones, dolomites, marlstones and even coals (Murascov and Mokin, 1979).

The Devonian sequence on Spitsbergen was deformed during the Svalbardian Orogeny in Upper Devonian times (Vogt, 1928; Friend and Moody-Stuart, 1972). A major left lateral strike-slip fault was possibly responsible for the displacement of the Hornsund Devonian towards the south (See below). Faults developed in a N-S direction, and folds are restricted to a number of N-S striking zones (probably older, reactivated lines of weakness), which reflect movements in the deeper basin. Within these zones a cleavage has been observed (Torsvik et al., 1985). An oblique stress field acted, causing folds to get a NNE-SSW-trend.

A more extensive theory was proposed by Harland and Wright (1979). They maintained to recognize three major sinistral fault zones with displacements in the order of 250-750 km. This implies that the Devonian on Spitsbergen was deposited in areas then much farther to the north, and subsequently whole areas were transported southwards during Upper Devonian times. This theory was recently discounted by Torsvik et al., (1985), who, based on palaeomagnetic evidence, concluded that Svalbard and the Barents Sea region have formed a stable platform since late Devonian/early Carboniferous time.

During the opening of the Atlantic Ocean in Mid-Eocene times, a dextral, transpressional regime acted. This deformational episode has been called the West Spitsbergen Orogeny.

2.4 NORWAY

The O.R.S. in Norway is mainly situated in two regions (Holtedahl, 1960). The largest region is in Western Norway, where sediments onland are preserved in four separate areas (Fig. 2.2). Those are Hornelen (Kolderup, 1915), Håsteinen (Kolderup, 1925), Kvamshesten (Kolderup, 1923) and Solund (Kolderup, 1926). Many islands to the west are also covered with "Devonian" sediments (Kolderup, 1916; Kolderup, 1927).

The other region containing "Devonian" sediments is situated farther to the NE (Fig. 2.2). The sediments crop out on Fosen and Storfosna (Vogt, 1924; Vogt, 1929; Richter, 1958; Siedlecka, 1975), Hitra (Reusch, 1916; Peacock, 1965; Siedlecka and Siedlecki, 1972), south of Smøla (Peacock, 1965) and on some islands northwest of Kristiansund (Bryhni, 1974). In addition there are some outcrops on the small islands NE of Frøya and Fosen (Vogt, 1929; Sæbøe, 1972).

Finally, sediments of "Devonian" age are found in the small Røragen Area E of Røros (Goldschmidt, 1913; Roberts, 1974). Sediments of Ludlovian-Downtonian age do also exist in the Oslo Region (Turner, 1974). Summaries of the Norwegian O.R.S. areas have been made by Nilsen (1973), Roberts (1983) and Steel et al. (1985).

The O.R.S. in Norway was deposited during a long time span. The age of the deposits ranges from possibly Wenlockian and Ludlovian in Middle and Upper Silurian to Upper Devonian. However, all sediments have been termed "Devonian" or "Downtonian". The region with outcrops in Western Norway is about 70x120 km, and is subdivided into smaller areas by different fault systems and structural features. The northern region between the mainland and the islands is more than 160 km long, but is nowhere broader than about 8 km. The maximum stratigraphical thickness of the O.R.S. in Norway is measured in the Hornelen Devonian, where it reaches more than 25 km (Bryhni, 1964). However, the vertical depth is probably not more than 8 km (Steel and Gloppen, 1980). In Western Norway there are suggestions for at least some of the areas (Hornelen and Kvamshesten) to be original structural basins seen by marginal facies interfingering with more distal facies (Bryhni and Skjerlie, 1975; Steel and Gloppen, 1980). Basin margins or marginal faults can nowhere be observed, because later faults cut into the deposits. The Solund Devonian is preserved over a much wider area than what could originally have been a restricted basin (Nilsen, 1968; Nilsen, 1973).

In the northern region no signs of basin margins are clear, and the deposits probably originally covered a much larger area (Ch. 7.2).

The O.R.S. in Norway rests unconformably on various igneous and metamorphic Precambrian and Caledonian rock complexes, and are mainly derived from these. In the western region palaeocurrent directions mostly point in northerly, westerly and southerly directions (Nilsen, 1968; Nilsen, 1973; Steel et al., 1977). In the Trøndelag and Nordmøre region palaeocurrents are generally directed either to the ENE or WSW, but there are large variations from place to place (Peacock, 1965; Siedlecka and Siedlecki, 1972; Siedlecka, 1975). Palaeocurrents directed both to the north and south were observed by the mentioned workers.

Tectonism played an important role during deposition of the Devonian in Western Norway. The concept of migrating basins was first introduced by Bryhni (1964), and later it was further developed by Steel and his co-workers (Steel et al., 1977; Steel and Gloppen, 1980). In the Hornelen Devonian renewed uplift during deposition led to more than 200 coarsening upward cycles (Bryhni, 1964; Steel et al., 1977; Steel and Gloppen, 1980). Both lateral and vertical variations in sedimentary facies exist. The most prominent lithologies are conglomerates, sandstones and siltstones. In Outer Trøndelag there are also minor occurrences of dolomite and calcite nodular horizons. The sedimentation was in a continental intramontane environment with deposition on alluvial fans, in braided and meandering rivers and in lakes (Steel et al., 1977; Steel and Aasheim, 1978). No proof has been found for marine influence. The calcareous horizons probably represent calcretes (caliches) developed on fan surfaces and as soil horizons in overbank areas. Dolomite was deposited on playas and in ephemeral lakes (Bryhni, 1974; Siedlecka, 1977).

Volcanism is recorded only in one of the Norwegian O.R.S. areas. In the

Solund Devonian there was extrusion of rhyolitic and trachytic lavas, probably in the early Middle Devonian (Furnes and Lippard, 1983). However, the age of the volcanics is still not exactly known. The geochemistry was suggested to be indicative of continental rift valley volcanism (Furnes and Lippard, 1983).

The O.R.S. in Norway has by a number of authors been considered as representing deposition during an extensional tectonic phase with NW-SE crustal stretching following the main Scandian Orogeny (Bryhni, 1964; Nilsen, 1968; Steel and Gloppen, 1980), but local, syn-depositional compression has also been suggested (Bryhni and Skjerlie, 1975; Steel and Gloppen, 1980). Two fault directions are dominant, one NE-SW and one NW-SE. The NE-ENE-trending faults in Western Norway and Trøndelag have been considered to be first order, dextral splays from a main sinistral mega-shear in the Atlantic region during the Caledonian Orogeny (Storetvedt, 1973; Ziegler, 1978; Kent and Opdyke, 1978; Van der Voo and Scotese, 1981). These splays were interpreted to have been active during deposition of the O.R.S., and in a transtensional tectonic regime structural basins developed (Steel and Gloppen, 1980).

Large-scale, sinistral movement along the Great Glen Fault in the Devonian and Early Carboniferous has recently been shown probably to be wrong (Smith and Watson, 1983; Irwing and Strong, 1984; Torsvik et al., 1985). Therefore dextral splays affecting sedimentation in Norway (Ziegler, 1978) are probably not real. The faults probably existed, but their strike-slip components (Steel and Gloppen, 1980) may have been neglible. The mechanisms behind the formation of migrating basins (Bryhni, 1964; Steel, 1976; Steel and Gloppen, 1980) are still not known.

During the Svalbardian contractile phase (Vogt, 1929) in the Upper Devonian (termed Røragian from Røragen by Roberts (1974) and Solundian from Solund by Sturt (1983)) the NE-ENE-trending faults developed into reverse faults and thrusts (Nilsen, 1968). The use of the term "Svalbardian Orogeny" for Western Norway and Trøndelag implies that these areas belonged to the same plate as Svalbard during deformation. This has possibly been the case (Torsvik et al., 1985), and the term will therefore be used later in this thesis. At this stage the sediments were deformed into open folds with axial trends varying from NE-SW through ENE-WSW to E-W. Synclines dominate with a general plunge to the ENE (Holtedahl, 1960).

In a new theory Hossack (1981) attemts to explain the "thrusts" in Western Norway as a flat lying, extensional, listric fault superimposed on a precursor Caledonian thrust fault. This theory could also explain slight folding, but not the large, tight folds in Håsteinen (Vetti, 1986). Neither does it explain how 25 km of sediments could be deposited in the Hornelen area (Steel, Siedlecka and Roberts, 1985) nor metamorphism and cleavage (Sturt et al., 1986). Reinvestigation during the summer of 1985 (Sturt et al., 1986), has provided evidence for eastwards thrusting of the Hornelen and Kvamshesten Devonian during deformation. These thrusts, which cut the unconformity, are not reactivated, earlier normal faults, as suggested by Nilsen (1968) in the Solund Devonian. The sequence of deformation was first folding, then thrusting and finally continued folding (Sturt et al., 1986; Torsvik et al., 1986).

The "Devonian" successions have been subjected to low grade metamorphism (Furnes and Lippard, 1983; Sturt et al., 1986; this thesis). This poses a serious problem in the interpretation of the deformation, as high pressures and temperatures (300-400^{0}C) must be associated. A possible major orogenesis post-dating deposition of the Devonian, perhaps involving major thrusting (Sturt, 1978; Sturt, 1983; Sturt, 1984), is further discussed by Sturt et al. (1986). Thrust-nappes on top of the now preserved "Devonian" could better explain the metamorphism than just burial below younger sediments. A succession more than 10 km thick would be needed on top of the "Devonian" to reach 300-400^{0}C.

In the Oslo Region there was sedimentation from Wenlock (marine) through Ludlov (continental) into the Downtonian without breaks (Turner, 1974; Worseley et al. 1983). Towards the NW the succession is increasingly thrusted and folded. This has been interpreted as a possible distal expression of the Scandian orogenic phase (Sturt, 1978). Decollement occurred as a result of gravitational spreading and/or compression from the NW (Bockelie and Nystuen, 1985). Others have mentioned the possibility that these structures are "Svalbardian" (Roberts, 1974; Sturt, 1978).

Sturt (1983) indicated that Variscan deformation may have influenced Norwegian O.R.S. deposits too.

2.5 SCOTLAND

In Scotland Devonian rocks are mainly preserved in two areas; within and S of the Midland Valley Graben and in the Orcadian Basin of NE-Scotland and the Shetlands (Anderton et al., 1979; Mykura, 1976). The sedimentary pile may reach substantial thicknesses, more than 10 km on Shetland, a little less in the Midland Valley. Sediments were shed into these basins from the surrounding uplifted margins. Marginal alluvial fans pass into braided rivers, meandering rivers and lacustrine sediments in the centres of the basins (Leeder, 1973). Calcareous mudstones and limestones regarded as lacustrine deposits of periodically drying lakes are found. The same interpretation has been put forward for the flagstones, which consist of 0,1-1 mm thick laminites of silt, micritic dolomite and limestone and some phoshate (Donovan, 1975).

Fining upward sequences of conglomerate, sandstone and red siltstone indicative of meandering rivers and fluvio-lacustrine regressive cycles are common (Leeder, 1973). Another typical feature is the cornstones composed of carbonate nodules and massive carbonate beds (Burgess, 1961). They are interpreted as ancient soil horizons. In the Upper Devonian there is evidence of temporary marine transgressions from the Hercynian Tethys to the S. Marine Middle Devonian reef limestones have been found in the Auk Oilfield east of the Midland Valley Graben (Anderton et al., 1979).

Bordering the Midland Valley are the Higland Boundary Fault to the north and the Southern Uplands Fault to the south. Both were probably active during deposition, and the Midland Valley was therefore a real graben (Anderton et al., 1979). The Orcadian Basin was a rapidly subsiding intermontane lake basin. In the Shetlands there were several smaller basins (Mykura, 1976).

Devonian rocks contain a large percentage of lava fragments. This reflects the extensive volcanism during the Lower and Middle Devonian.

Calc-alkaline, andesitic, rhyolitic and basaltic lavas, ignimbrites, tuffs and agglomerates are common in the Midland Valley. Swarms of calc-alkaline granites and granodiorites termed the "Newer Granites" were intruded in the Midland Valley (Anderton et al., 1979).

During Late Silurian-Early Devonian times a deformational episode took place in the Southern Uplands (Anderton et al., 1979). Another tectonic episode in Middle Devonian times threw the Lower Old Red Sandstone of the Midland Valley into a series of open folds. Middle Devonian sediments are lacking, but in the Upper Devonian the Midland Valley was again a deep sedimentary basin. In the Orcadian/Shetland Basin mainly Middle Devonian sediments were deposited. In the Caithness area Upper Devonian rocks rest unconformably on Middle Devonian sediments (Mykura, 1976). Chemical analysis of the lavas in the Midland Valley and in the Orcadian Basin/Shetlands indicate that they are related to subduction (Thirlwall, 1981). Thirlwall (1981) suggested that subduction during the Caledonian Orogeny stopped during the Lower Devonian in the Midland Valley area, but continued into the Middle Devonian in the Orcadian Basin/Shetlands. Closure of Iapetus in this region was perhaps as late as mid-Devonian. Polyphase deformation in the Shetland O.R.S. was supposed to support this idea (Thirlwall, 1981) (Ch.11.1).

On Shetland the sediments are locally strongly folded and slaty cleavage may be developed. In these areas metamorphism reaches lower greenschist facies. Fold axes generally strike NE-SW, and faults trend N-S (Mykura, 1976). The deformation and metamorphism have earlier by a number of authors been ascribed to large scale left-lateral movement along the Great Glen Fault System, but recently this movement has been shown probably not to be real (Ch. 2.4). However, smaller scale, vertical and lateral movements along faults occurred (Mykura, 1976).

2.6 COMPARISON BETWEEN THE O.R.S. AREAS

According to most previous authors, sedimentation during Devonian times in the North Atlantic Region took place on and around the margins of the Old

Red continental landmass. The distribution of sedimentary facies indicates that Eastern Greenland and Western Norway were situated in the centre of this landmass, while Spitsbergen and Scotland were closer to the margins (Fig. 2.1). The centre was characterized by terrigenous clastics representing alluvial fans, rivers, lakes and playas. In Scotland and on Spitsbergen the situation was similar, but shorter periods with marine deposition occurred during transgressions (See below). Everywhere sedimentation was rapid due to actively subsiding basins. There is a spread of sedimentary ages from Upper Silurian throughout the Devonian, but no clear trend can be detected.

Both in Scotland, Eastern Greenland and on Spitsbergen the sedimentary basins had a large areal extent. In Norway some O.R.S. areas were apparantly rather restricted, structural basins (Røragen, Kvamshesten, Hornelen), while other deposits covered extensive areas (Solund, Outer Trøndelag). In Scotland very large-scale, syn-depositional faults have been identified, e.g. the Highland Boundary Fault. In Norway no syn-sedimentary faults of this magnitude have been identified, only inferred.

On Spitsbergen Devonian magmatic activity has not been detected, but in Scotland and Eastern Greenland volcanicity was widespread, and large plutons were intruded. Also in the Solund Devonian volcanicity has been observed. Lavas from Solund have been interpreted to be related to continental rift valley volcanism. In Scotland lavas have a geochemistry indicative of subduction, which may have taken place until the Middle Devonian. This latter theory does not fit the generally accepted model of a huge, Old Red Continent existing throughout the Devonian (Ch.11.1).

Both in Norway and Scotland evidence is found suggesting a deformational episode in Late Silurian-Early Devonian times (Anderton et al. 1979; Siedlecka, 1975). During Middle Devonian times deformational episodes have been documented in Eastern Greenland, Scotland and on Spitsbergen. The Upper Devonian deformation acted everwhere. This event was responsible for faulting and even overthrusting in Eastern Greenland and Norway.

CHAPTER 3

REGIONAL GEOLOGY, BASEMENT AND CONTACT RELATIONSHIPS

3 REGIONAL GEOLOGY, BASEMENT AND CONTACT RELATIONSHIPS

3.1 REGIONAL GEOLOGY

The "Devonian" rocks in the Trøndelag Region are situated within the Bergen-Namsos Gneiss area (Oftedal, 1981). The rocks of the Precambrian basement are considered to be about 1700 m.y. old both with respect to age, metamorphism and deformation, but later deformational phases are detected.

The Precambrian rocks between Trondheimsfjord and Romsdal have been subdivided into 5 groups (Oftedahl, 1981). The tectonically lowermost is the Frei Group, which consists of supracrustal rocks like augengneiss, massive gneiss, granitic gneiss, pelitic migmatites, quartzite and marble. Lenses of eclogite and dolerite are found within these rocks. Then comes the Rausand Group, consisting of reddish granitic gneisses with dolerite sills and the Tingvoll Group which appears to underlie Hitra to the SE. It is made up of homogeneous gneiss, augengneiss, quartzite, amphibolite and marble. Above follows the Sjuråsen Group with mica schist and amphibolite, and finally the Gangåsvann Group with amphibolitic mica schist and phyllite. The area is tightly folded, with axial planes striking NE-SW. The same rocks also crop out on Frøya N of Hitra. This island is mainly made up of granitic and granodioritic migmatitic gneisses.

Southeast of the Precambrian basement one passes into southeastwards overthrusted Caledonian rocks of Precambrian to Silurian age. A summary on Caledonian deformation in Norway was made by Roberts and Sturt (1980). The rocks are tightly folded in dominantly one northwestern syncline and one southeastern anticline. The strike is NE-SW. In the past the stratigraphy has been described in many different ways due to difficulties in understanding the thrusting. According to Roberts and Wolff (1981), the northwestern area around Trondheim is composed of a Lower Nappe Complex and an Upper Nappe Complex overlying the parautochthonous and autochthonous basement. The lower complex is composed of the Leksdal, Skjøtingen and Levanger Nappes, the upper complex of the Gula and Støren Nappes.

The Lower Nappe Complex contains rocks ranging from Precambrian to Ordovician in age. The rocks are from bottom to top quartz schist, augengneiss, deformed Rapakivi Granite, gabbro, anorthosite, amphibolite and hornblende mica schist.

The Gula Group of the Gula Nappe predominantly consists of metasandstones, phyllites, mica schists and amphibolites. The central part of the Gula Group has been subjected to strong metamorphism, generally regarded as Barrovian. Originally the rocks of this group were deposited in an epicontinental environment. Above this comes the Støren Nappe with four different groups. They are all of Cambro-Silurian age. The lowermost one is composed of greenstone, then, above a thrust, come different metasediments and metavolcanics. The Gula Group and the lowermost Støren Group of the Støren Nappe are intruded by gabbros and trondhjemites.

On Smøla SW of Hitra there is a sequence of Lower-Middle Ordovician conglomerates, limestones and metabasalts, which are thought to be the continuation of the Støren Nappe. It is cut by a later Caledonian intrusive complex ranging in composition from ultramafic to granodioritic.

3.2 BASEMENT ON HITRA

On Hitra one can generally observe the same rock types as on Smøla. In the NW there is a sequence of gneisses, mica schists, amphibolites, marbles and granodiorites of Mid-Ordovician age (Kollung, 1964). The rocks have been correlated with similar rocks in the Trondheim Region and on Smøla, and probably have the same origin. The largest part of Hitra is composed of a dioritic rock. This is also probably of Mid-Ordovician age (Kollung, 1964). Later there has been intrusion of large masses of granite and quartz diorite on SW-Hitra. The latest generation of intrusives is a porphyric diorite and the ultramafic rocks on Helgebostadøy.

3.3 ROCKS UNDERLYING "DEVONIAN" SEDIMENTS AND CONTACT RELATIONSHIPS

3.3.1 Introduction

The "Devonian" sediments on Hitra are everywhere underlain by igneous rocks or by metamorphic rocks of igneous origin. The intrusive body on which the Hitra Group rests has a composition that varies from granitic through granodioritic to dioritic. The diorite is composed of plagioclase, amphibole, chlorite, biotite, quartz and sphene (Fig. 3.1), and is medium to coarse grained. Xenoliths in the diorite have a similar composition, but the grain size is finer. There are gradual transitions into granodiorite and granite, which have larger fractions of quartz and K-feldspar. The feldspars exhibit a varying degree of recrystallization.

Diorite is the predominant lithology of the undeformed basement, therefore the field term "diorite" will be used throughout. Time was not sufficient to allow mapping out of compositional variations. Only at specific localities was the composition examined in detail. Sometimes the diorite contains up to 50 % xenoliths which often have a rounded shape. In places with strong regolith development on top of the diorite it could be difficult to see the difference between dioritic basement and basal, sedimentary breccia (Figs. 3.2 and 3.3). Neither the diorite nor the xenoliths are much deformed. For further discussion, see below.

The boundary between the sediments and the diorite to the NW is defined by an unconformity. Prior to deposition the diorite was deeply eroded and denudated.

3.3.2 The Northern Margin

The pre-sedimentary terrain was characterized by an undulating surface with valleys up to 100 m deep and 1 km wide (Ch. 6.2.1 and Plates 1A and 1B). The Basal Conglomerate of the Hitra Group predominantly accumulated in these valleys on top of the regolith. On the ancient hills it is easy to observe the exact diorite-sediment contact, because finer sediments, higher in the succession, lie directly on the intrusive rocks. Sandstone-filled cracks in the old diorite surface have been found on such hills, for example at the eastern end of Balsneshusvann (611650 23300). At this locality there is what must be an ancient hill piercing through the sedimentary pile. The size of the exposure is about 75x40 m. On the southern side there is an unconformity with undulating relief less than 1 m high. In the topographic depressions 10-30 cm of conglomerate occur. Upwards it passes into horizontal bedded and cross-bedded sandstone (Fig. 3.4). Fig. 3.5 shows a sand-filled crack about 0.5 m below the unconformity. In this place the regolith is about 1 m thick.

20 m to the west of this locality, on the northern side of the ancient hill in the westernmost part of the outcrop, a 3 m thick regolith is preserved. The unconformity is also easily observable there. 5 m to the west there is another small islolated outcrop of regolith with sand-filled cracks. Generally the regolith is best developed on the northern side of the palaeohill/palaeocliff.

Sand-filled cracks were also observed at a site just to the S of Strandavannet (116800 35600). The classical locality where an unconformity has been demonstrated is in the W, just to the E of the mouth of Aunåa (610500 19900) (Reuch, 1914; Peacock, 1965; Siedlecka and Siedlecki, 1972). For further discussion, see Chapter 4.

Everywhere along the northwestern margin there appears to be an unconformity between basement and the "Devonian" sediments. A regolith is commonly developed. The contact between basement and sediments is best seen on top of palaeohills. This is because erosion/denudation was more intense there. As a regolith formed, it was rapidly eroded away. In palaeovalleys, on the other hand, there was little erosion, and more time was available for formation of a thicker regolith. As already mentioned, it is difficult to discern basement from basal, sedimentary breccia in those areas.

3.3.3 The Southern Margin

The unconformity can be observed only at one locality along the southern margin of the "Devonian", on a small promontory to the NE of Bekkvikholmen (610800 24250). The locality is accessible at low tide only. The diorite/regolith is overlain by 1-5 m of conglomerate/basal breccia (Fig.3.6), which again is succeded by finer grained sediments. The diorite is cut through by a number of epidotized cracks and microfaults, but is otherwise little deformed. On Bekkvikholmen to the S (610600 24150) the diorite is strongly brecciated by a ENE trending fault. Fragments are generally less than 15 cm across.

The existence of an intrusion (Ch. 10), and the quite different surficial appearance of the sediments at this locality led Siedlecka and Siedlecki (1972) to conclude that they were of Ordovician age. Closer examination under the microscope has revealed that they belong to Member B (Ch. 4). Further W there are two more localities with similar looking sediments, where it is possible to see the primary erosional contact on top (Fig. 3.7). The structural and sedimentological features correspond closely with the surrounding "Devonian" trends (Chs. 6 and 8). These points provide evidence strong enough to conclude that the sediments belong to the O.R.S. succession.

Also on Selnes (610800 25200) Siedlecka and Siedlecki (1972) described Cambro-Silurian deposits, but the structural and sedimentological similarities with nearby "Devonian" rocks suggest that also they belong to the O.R.S. succession. This was pointed out by Roberts (1981).

At all other localities where it is possible to see basement against "Devonian" the contact is a fault/shear zone. Probable attitudes of faults during their formation will be discussed in Ch. 8. If one starts in the SW, the succession is faulted against an intrusion of trondhjemitic, tonalitic and quartz dioritic rocks (610500 19900-611000 20000). Little deformation has taken place in this. The next place to observe basement is on Steinbitholmen (610250 23600). The southern part of the island is composed of rather undeformed diorite, with a number of granitic intrusions. A 10 cm thick, vertical, cataclastic shear zone, trending

ENE-WSW, separates it from Member B sediments to the N (Ch. 4).

The same relationship occurs on Kalvhaugterna (610350 24000), with diorite intruded by granitic veins and pegmatites to the S separated from the Vollan Formation (Ch. 4) to the N by a 2 m thick, vertical cataclasite (Fig. 3.8). Its present attitude is ENE-WSW. The next place to observe diorite was on Langnes (611250 26000-611250 26500), where there are two exposures separated by a right lateral, sub-vertical shear zone at least 4 m thick and trending NW-SE. At the westernmost outcrop the northwestern boundary is below sealevel, the exposure being at a skerry. The easternmost outcrop is fault bounded both to the NW and N by two separate faults. Thin, 1-2 cm thick mylonites have been produced along the sub-vertical to southwards dipping faults. The diorite is a grey, medium grained rock. The Vollan Formation lies on the northern side of the faults. Neither on Langnes nor on Kalvhaugterna is the diorite much deformed.

The second last locality where basement can be seen in contact with the "Devonian" is on an approximately 2 km long coast section running westwards from Kallarnes (614700 34600). A NE-SW trending sinistral fault has offset the main shear zone between Hitra and Jøssenøya to the N. The displacement is probably > 200 m (Plates 1A and 1B), but the fault is nowhere exposed. The basement resembles very closely the rocks which are found on Jøssenøya and the other small islands to the south of Hitra (Fig.3.9). It is a strongly deformed dioritic rock, which, prior to , and perhaps subsequent with deformation, was intruded by a number of granitic and pegmatitic veins and dikes. To the N one passes into metasediments of sandstone and siltstone, but the exact boundary can not be seen because of strong deformation. Deformation decreases over a distance of 150 m to the N of the contact. Both basement and nearby sediments are in places mylonitized (Fig. 3.10). The position of the sheared boundary was picked from thin section studies (Plates 1A and 1B). Close to the boundary both basement and sediments show a strong cataclastic texture (Ch. 8). The age relationship between ductile and brittle textures will be discussed in Ch. 11.

Usually it is difficult to recognize the protolith, but there are zones parallel to foliation where it is less deformed. There one can clearly see the original diorite, which is intruded by a pink pegmatite. Siedlecka and Siedlecki (1972) claimed these rocks partly to be meta-diorite and partly to be Cambro-Silurian meta-sediments. The present author could not find evidence for the existence of Cambro-Silurian meta-sediments. The meta-sediments described by Siedlecka and Siedlecki (1972) are an integral part of the "Devonian" succession.

A general impression is that pegmatite-rich areas have been more competent than the surrounding rocks during deformation. This is not as one should expect, as a diorite should behave more competently than a granite during deformation. It can either be a result of late intrusion of granite bodies during deformation, or perhaps the diorite was already slightly altered/deformed prior to granite intrusion and subsequent strong deformation. Both possibilities have to be considered, since there are varying degrees of deformation of the intrusions. There are boudins of granite pegmatite with little deformation about 15x7.5 m in size which occur within strongly sheared diorite.

It is to be hoped that U/Pb-radiometric dating can be done on zircons from the intrusions. Such an analysis was not carried out in this thesis because of time limitations, but it could tell much about timing of both intrusion and deformation. Possibly was the intrusion contemporaneous with the intrusion of pegmatitic dikes in the Kristiansund area (Pidgeon and Råheim, 1972; Ch. 10).

From thin sections it can be demonstrated that the basement is cut through by large numbers of cracks and microfaults. These are filled with epidote, calcite and quartz cement. This is also the case for nearby sediments. The green colour comes from the large amount of secondary epidote, which sometimes occur as big concentrations of crystals (Ch. 9.2.2.3).

Finally, in Grindvik (619800 40500), there is an E-W running shear zone at least 10 m wide separating the Vollan Formation from the diorite to the E. Apparantly there has been strong brittle shear, but neither the sediments nor the diorite along it are much deformed. The shear zone dips about 75 degrees to the south.

An area of basement rocks occurring within the sedimentary succession SE of Litlvann will be discussed in Ch. 8.9.2.

CHAPTER 4

THE HITRA GROUP

4 THE HITRA GROUP

4.1 GENERAL NAMES AND SUBDIVISIONS

The Hitra Group is identical with what has earlier been termed "The Hitra Beds" (Peacock, 1965) and "The Hitra Formation" (Siedlecka and Siedlecki, 1972), and includes all O.R.S. deposits on Hitra. In agreement with Siedlecka it was decided still to make use of the name "Hitra". The succession was subdivided into three formations, which are mappable throughout the area. They are from bottom to top the Aune Formation, the Vollan Formation and the Balsnes Formation. There are clear lithological differences between them.

From Table 4.1 it can be seen that the names are the same as Peacock (1965) used. However, according to the new terminology it is not correct to define them as groups. He further divided the three groups into twelve formations, of which some of them only constitute a few beds. A total re-subdivision into members was therefore made. The reason for not using Siedlecka and Siedleckis' (1972) subdivision is that it is not possible to distinguish their Members C and D as one goes eastwards. They are therefore not mappable units. In addition, a further subdivision was needed in the westernmost area for the sedimentary environment descriptions.

4.2 TYPE SECTION

Along the western coast from Balsnes (610500 23000) to the mouth of Aunåa (610500 19900).

FORMATIONS	MEMBERS	THICKNESS	LITHOLOGY	Earlier subdivisions	
				Peacock (1965)	Siedlecka & Siedlecki (1972)
Balsnes Formation	L	ca. 200 m	Conglomerate	Balsnes Congl. Form. m	Member E
Vollan Formation	K	ca. 250 m	Alternating sandstones and mudstones	Vollan Group Form. d-l	Member D
	J	48 m	Mudstone		
	I	127 m	Alternating conglomerates and sandstones		
	H	12 m	Mudstone		
	G	79 m	Alternating sandstones and conglomerates		
	F	157 m	Alternating sandstones and mudstones		
	E	40 m	Sandstone		
	D	14 m	Alternating mudstones and sandstones		
	C	ca. 240 m	Conglomeratic sandstones		Member C
Aune Formation	B	53-163 m	Alternating red and grey sandstones and siltstones	Aune Group Form. a-c	Member B
	A	43-114 m	Basal breccia/conglomerate		Member A

Table 4.1 The stratigraphy of the Hitra Group.

4.3 THICKNESS

The maximum thickness of the succession is measured in the westernmost part of the area. It has partially been obtained from logs and partially calculated from the maps using amount of dip. A minimum of 1350 m was found. Eastwards there are variations, but a general decrease in thickness is recorded. In Grindvik (119800 40500) there is not more than 150-200 m preserved.

4.4 LITHOLOGY

The sedimentary pile commences with a basal breccia, which passes upwards into red mudstones, sandstones, sandstones and mudstones and finally into conglomerate.

4.5 BOUNDARIES

The lower boundary of the Hitra Group is an unconformity upon crystalline basement (Ch. 3.3). Possible younger rocks on top of the "Devonian" are nowhere preserved.

4.6 DISTRIBUTION

The group occupies all the area along the coast of Hitra from Aunåa in the SW to Grindvik in the NE, and reaches between one and two km northwestwards onto the island. The length of the area is about 23 km.

4.7 GEOLOGICAL AGE

The age of the Hitra Group is still not exactly known. A Downtonian age was suggested from fossils found by Reusch (1914). Størmer (1935) identified them as *Dictyocaris slimoni* Salter and *Hughmilleria* sp.. Recently it has been documented that these fossils existed both during the

Wenlockian, Ludlovian and Lower Devonian. No fossils were found during field work on this project. Neither were spores nor plant remains observed. Also Allen (1976) searched the Hitra Group for spores with a negative result. However, bioturbation is common in Member B (Chs. 4.8.7.5 and 6.2.2). The age problem is further discussed in Chapter 7.1.

4.8 AUNE FORMATION

4.8.1 General

This formation was originally described by Peacock (1965) as the Aune Group (Ch. 4.1), and constitutes the lowermost deposits of the Hitra Group. Lithology, boundaries and distribution will be discussed under the member descriptions.

4.8.2 Name

The name is taken from the farm Aune, which is situated just to the E of Aunåa in the westernmost part of the area (610900 20200).

4.8.3 Type Section

The small peninsula just to the E of the mouth of Aunåa. The locality was first described by Reusch (1914), and has also been used by later authors. The section is about 65 m thick.

4.8.4 Thickness

The formation has a variable thickness ranging from more than 350 m north of Litlvannet (616200 34750) to being absent other places.

4.8.5 Subdivisions

The formation is subdivided into two members, the lowermost Member A, and Member B. Both are mappable throughout the area of investigation. The reason for classifying the two members as a separate formation is found in their similar sandstone composition.

4.8.6 Member A (Basal breccia/conglomerate)

4.8.6.1 Type Section

Peninsula and coast just to the E of the mouth of Aunåa. The section is about 45 m.

4.8.6.2 Thickness

The thickness varies from zero to a maximum of 175 m NE of Badstuvik (616900 37700), but normally, when it is present, it is not more than 100 m thick.

4.8.6.3 Lithology and Petrography

This member is dominated by poorly sorted material. Maximum boulder size is about 1.5 m, and there is a complete range down to coarse grained sandstone. The largest boulders are found near the bottom of the deposit. Upwards there is a gradual decrease in size. The largest cobbles on top of the member are about 15 cm in diameter. There is also a decrease in amount of boulders and cobbles. At the same time the fragments become better rounded upwards. The fragments near the bottom are subangular to subrounded (Fig. 4.1), while fragments higher up are subrounded to rounded. The breccia is generally matrix supported.

Grain size and degree of rounding varies throughout the area. However, composition of the member always reflects composition of local basement.

This may vary from dioritic through granodioritic to granitic. Also the sandy matrix reflects the composition of local basement. There is an increasing amount of exotic rock fragments upwards in the deposit.

4.8.6.4 Boundaries

The lower boundary of Member A is also the lowermost boundary of the Hitra Group, and is an unconformity. The unconformity was described in detail in Chapter 3.3, so this will not be repeated here. The upper boundary of the member is a rapid gradation into Member B. From cobbles of about 15 cm in diameter there is a 1 m thick transition zone into very coarse sandstones with scattered pebbles and vaguely defined conglomeratic horizons.

4.8.6.5 Distribution

As mentioned in Chapter 3.3, outcrops of this member are scattered along the margins of the area. Traced from the west, it is found from Aunåa (610900 20200) to Balsneshusvannet (611250 22800) and from Balsneslangvannet (612250 24700) to the W of Aksethusvannet (213700 28200). Then it reappears 1 km east of Terningvannet (615500 33300), and continues to the south of Storvågen (618000 36900). A thin development is also found to the southwest of Litlvågen (619000 38500). Along the southern margin it is only found on the promontory northeast of Bekkvikholmen (610800 24250).

4.8.7 Member B

4.8.7.1 Type Section

A typical section through the lower part of the member is seen at the headland E of the mouth of Aunåa. About 20 m can be observed.

4.8.7.2 Thickness

There is a range in thickness from zero to a maximum of about 190 m N of Litlvannet.

4.8.7.3 Lithology and Petrography

There is a range in grain sizes from coarse sandstone to mudstone. Mudstones are commonly dark red, but may be dark green, as along the southern margin of the area. Sandstones are grey, reddish grey or greenish grey. Usually Member B starts with a coarse, 2-50 m thick sandstone. Then there is a gradual increase in number and thickness of mudstone beds. At the type locality this change starts 15 m above the base.

The sandstones are poorly sorted, and the grains are generally subangular. A typical composition is seen from Table 4.2 (Number 18703). Carbonate is the dominating cement, and may constitute more than 15 volume percent of the rock. Other cements are silica, fine grained, white mica and epidote. Feldspars and biotites are retrograded with production of white mica, possibly sericite and chlorite, respectively. Small epidote-filled veins cut through the sandstones. The mudstones are generally too fine grained for compositional determination. There are laminae of what probably originally were clay minerals, and coarser laminae rich in quartz and feldspar grains.

A typical feature of Member B is that it is very rich in calcareous concretions (Fig. 4.2). Their length may be up to 10 cm, while the thickness is normally not more than 3 cm. Usually they are arranged in well defined horizons. Siedlecka (1977) examined the concretions in detail (Ch.6.2.2), so they will not be further discussed here.

The composition of sandstones of Member B is much the same as the matrix composition of Member A. The typical appearance of Member B is shown in Figs. 4.3 and 4.4.

Locality	Balsnesaune	NW of Olsvik	Akset	Oddskjeret	Selnes	Kalvhauterna	SW of Litlv.	Fuggelås	Grindvik	Badstuvik
Sample number	18703	20264	18701	18705	18968	19215	20346	20768	20345	20766
Grid reference	611000/ 20250	612000/ 26000	613650/ 30500	609900/ 22150	611000/ 25500	610350/ 24000	615500/ 34450	617100/ 36200	119850/ 40450	116800/ 37600
Rock/Mineral Type	%	%	%	%	%	%	%	%	%	%
K-feldspar	2.0	5.5	4.0	15.0	9.0	3.5	19.5	14.5	1.0	6.5
Plagioclase	8.0	15.0	46.5	23.0	16.5	25.5	46.0	40.0	25.0	36.0
Quartz	24.0	28.5	16.0	18.5	24.0	30.5	8.5	23.5	44.5	28.0
Muscovite	0.5	1.5	0.5	2.0	1.0	0.5			1.0	0.5
Chlorite	2.5	3.0	3.0		7.0	1.0	3.0	2.0	4.0	1.0
Epidote	5.0	6.0	14.5	4.5	8.0	4.5	2.0	12.0	14.0	4.5
Accessories			0.5	0.5	0.5	1.0		0.5	1.5	0.5
Biotite	9.5	0.5								
Erts	2.5	1.5			1.5					0.5
Amphibole		0.5								0.5
Matrix		16.0	13.5	9.0	9.5	15.0	7.0	3.0	4.5	10.5
Calcite cement	14.5	6.0		0.5		2.5				5.5
Siltstone		1.5						0.5	0.5	1.0
Chert		4.0		5.0	7.5	3.5	1.0	1.0	1.0	
Porphyry		4.0		6.0		1.5	8.0	1.0	0.5	3.0
Gabbro				4.5		8.5				
Diabas		0.5		4.5	6.0	0.5	1.0			
Basalt		0.5	1.5	2.0	5.0			2.0		1.0
Andesite		1.5		2.5			3.0			0.5
Schist/Gneiss		3.5		2.5	2.5	1.5	0.5		1.0	0.5
Granite/Diorite		0.5			2.5		0.5		1.5	
Undiff. rock fragments	21.0									

Table 4.2 Petrography of sandstones. Percentages are obtained from 200 counts in each thin section. Notice that sample nr. 18703 is from the Aune Formation and sample nr. 20264 is from the Balsnes Formation. All others are from the Vollan Formation.

4.8.7.4 Boundaries

The lower boundary has been described in Ch. 4.8.6.4, but where Member A is absent, Member B lies directly on basement. Then there is the same kind of unconformity as described in Ch. 3.3. The upper boundary appears to be either gradational or erosional. In the cliffs between Grunnvann (612050 22400) and Balsneshusvann there is a gradual change from Member B into Member C over a distance of about 10 m. In this zone, some grey mudstone beds 10-30 cm thick lie between light reddish and greyish sandstone beds.

An erosional boundary was found on the coastline just to the north of Steinbitholmen (Ch. 3.3.3). Member C has eroded a relief of about 30 cm down into Member B (Fig. 3.7). Above the contact there are scattered mudflakes with signs of plastic deformation. The biggest clast observed was about 40 cm across.

4.8.7.5 Bioturbation

This member is the only one that shows clear bioturbation. In places the sediment is so strongly bioturbated that it is difficult to discern bedding. The burrows are usually filled with grey sandstone or siltstone, but may also be filled with early calcite cement (Fig. 4.5) (Siedlecka, 1977). Usually they are 2-3 cm across, have an irregular outline and are lined with a film of fine, black silt/clay (Fig. 4.5). Burrowing has been most thorough in the mudstone beds.

4.8.7.6 Distribution

Usually Member B is found in close association with Member A. It is present from the mouth of Aunåa to Balsneshusvann. Then there is a small outcrop NE of Hamnalangvann (612100 26250) and two outcrops to the N and NE of Aksethusvann (613250 28900). The largest area with Member B is from about 1 km E of Terningvann to the SE of Storvågen (615000 33300-617500 36600). A thin development was also found to the E of Litlvågen.

Along the southern margin of the area Member B is described on the promontory to the N of Bekkvikholmen, on the island to the W of this, on a narrow strip along the coast between these two and on the coast to the N of Steinbitholmen. In Grindvik there is a 4 m thick development of Member B which appears to lie within the Vollan Formation. Only about 5 m of the "Vollan Formation" can be seen below. This sandstone is reddish, and resembles the sandstones found in the transition zone between Members B and C at Grunnvann (Ch. 4.8.7.4). Therefore this is probably a similar transition zone. On top of Member B there is an erosional contact. Sediments above clearly belong to the Vollan Formation.

4.9 VOLLAN FORMATION

4.9.1 General

The formation was first described by Peacock (1965) as the Vollan Group (Ch. 4.1), and constitutes most of the Hitra Group succession. It is analogous to Siedlecka and Siedleckis' (1972) Members C and D.

4.9.2 Name

The name is taken from the farm Vollan (610750 22150) on the coast NW of Balsnes.

4.9.3 Type Section

The formation is defined along the coast from Aspvikodden (610700 20250) to Balsnes. There are nearly continuous exposures throughout the section. Type sections for the different members are all along the same stretch of the coast.

4.9.4 Thickness

The maximum thickness of the formation is measured along the western coast where it is about 960 m. Variation eastwards is partly a result of the shape of the synclines (Ch.8), but in Grindvik the actual thickness is not more than about 70 m.

4.9.5 Lithology and Petrography of the Formation and its Members

The formation is built of a variety of mudstones, sandstones, conglomeratic sandstones and conglomerates. The coarser fractions are poorly sorted. Sandstones display a low degree of rounding. Composition of sandstones has been checked at eight different localities along the length of the area (Table 4.2). As can be seen, there is no clear lateral trend with respect to lithological changes. Neither are there significant vertical lithological changes in the sandstone members. However, there are large differences from place to place. As for sandstones, it is difficult to see clear lithological differences between mudstones. Colour of sandstones is light grey to greenish grey, while mudstones are dark grey to black.

Mudstones are generally too fine grained for compositional determination, but there are laminae of what probably origianlly were clay minerals and oxides and coarser laminae rich in quartz and feldspar grains (Fig. 4.6). The sandstones are rich in detrital quartz, feldspar and rock fragments. Matrix may constitute up to 16 % of the total rock volume. In addition there are often large amounts of secondary epidote and fine grained, white mica, possibly sericite. The most common minerals occurring in the matrix are epidote, quartz, white mica, chlorite and calcite. It seems likely that most of this material has been derived from feldspars. The feldspars exhibit a varying degree of alteration from nearly unaltered, to hardly recognizable, almost completely recrystallized fragments. Fig. 4.7 shows the typical appearance of a sandstone from the Vollan Formation.

A lot of different sand-sized rock fragments were observed, the most common are of intermediate porphyry, chert, gneiss, granite, basalt, schist and mudstone. The most common composition of pebbles and cobbles is quartz, granodiorite, granite, pegmatite and gabbro. In addition there are minor amounts of metasandstone, welded tuff, jasper, pyroxenite and fine grained, intermediate volcanics. Pebbles and cobbles are well rounded. Mudstone clasts and chips are scattered throughout the sandstones and conglomerates. Nothing more will be said about the petrography here. Detailed studies were carried out by Peacock (1965) and Siedlecka and Siedlecki (1972).

4.9.6 Subdivision

The formation is subdivided into Members C to K, which are all mappable along the coast. The subdivision is based on lithological differences.

4.9.7 Boundaries of the Formation and its Members

The lower boundary of the formation was described in Ch. 4.8.7.4. The Vollan Formation may also lie directly on basement where the Aune Formation is not present. This is so for Member C (e.g. between Balsneshusvann and Balsneslangvann, Member F (west of Terningvann) and Member K (eastern part of the area). The unconformity is described in Ch. 3.3.2.

The boundary beween Members C and D encompasses an abrubt depositional change from very coarse grained, trough cross-bedded sandstones (Member C) to greyish black, homogeneous mudstones of Member D. The best place to visit the contact is at the westernmost coast of Furuholmen.

The only place where the boundary between Members D and E can be seen is at the southwestern coast of Furuholmen. Member D is composed of dark mudstones. Above a planar, erosive surface comes coarse grained, tabular and trough cross-bedded sandstones of Member E.

The boundary between Members E and F is nowhere exposed, but probably represents a change in sedimentary environment without any major break

(Chs. 6.3.1 and 6.3.3).

The boundary between Members F and G is sharp and erosional, and represents a break in deposition. Above the erosional surface one finds conglomerates and conglomeratic sandstones with a large content of mudclasts and calcareous concretions. The mudclasts have a maximum size of 1x0.5 m, and show few signs of plastic deformation. Concretions reach a size of 10x60 cm, and have all different orientations, which show that they have been eroded from the underlying sediments. Similar looking concretions were found in mudstones and sandstones just below the erosional surface (tectonic rotation of the concretions will be described in Ch. 8.3). It is difficult to estimate the time span this hiatus represents, and how much sediment that has been eroded away. However, a certain amount of time is needed for some consolidation of the mudstones and formation of calcareous concretions. The boundary is only seen at the coast SE of Furuholmen (610800 21650).

On top of Member G there is a sharp, flat, depositional transition from very coarse sandstones to relatively massive, dark, sandy siltstones. The transition can only be observed at the coast southeast of Furuholomen.

The upper boundary of Member H is a sharp, undulating, erosive surface, with coarse grained sandstones of Member I above. It can be visited at the coast northwest of Balsnes.

The transition between Members I and J is more indistinct. There is a gradation over about 10 cm into parallel laminated, fine and coarse grained sandstone beds of Member J.

The upper boundary of Member J is sharp and erosional, and represents the base of a channel or a channel complex at least 12 m deep, which has incised down into Member J. Above the contact there are very coarse, conglomeratic sandstones with plastically deformed mudstone blocks up to 1x7 m across (Ch. 6.3.7). The transition can only be seen at the coast NW of Balsnes.

The upper boundary of the Vollan Formation can be observed only at three localities. In the southeasternmost part of the area at the coast in Grindvik, there is an erosional boundary to the Balsnes Formation. The Vollan Formation is probably only about 70 m thick at this place. Going from Member B, as described in Ch. 4.8.7.6, there is a sequence of coarse sandstones, which gradually passes into conglomerate. This conglomerate wedge is about 20 m thick. Then there is a gradual transition into coarse to fine grained sandstones with dark coloured mudstone beds. Finally, above a sharp erosional contact lies the Balsnes Formation with coarse conglomerates.

100 m to the northwest of the farm Akset, in the slope of the hill, there is a gradual transition from the Vollan Formation into the Balsnes Formation (613500 29700). Over a distance of about 10 m one passes from sandstone with mudstone beds into sandstone with conglomeratic horizons, and finally into conglomerate, which becomes coarser upwards. Strike and dip of bedding was unchanged across the transition zone. A similar transition was observed in the bend of the road just to the SE of the farm houses.

Nowhere else can the direct transition between the two formations be seen, but generally there is a tendency for sandstones of the Vollan Formation to become conglomeratic as one approaches the boundary. This is especially the case along the southern margin. From Badstuvik to Grindvik it was difficult to place a precise boundary between Member K and the Balsnes Formation because of an apparant gradual transition between them. The lithology of Member K reflects this by containing a larger number of pebbles which normally only occur in the Balsnes Formation, e.g. jasper. On the northern side of the main syncline the transition appears to be more abrubt, but east of Strandavannet Member K interfingers with the Balsnes Formation (Plate 1B).

4.9.8 Distribution of the Formation and its Members

The formation can be traced all along the area from Aspvikodden to Grindvik, but this is not so for the different members (Plates 1A and 1B).

Member C starts at the fault W of Aspvikodden, and can be traced to about 600 m to the west of Aksethusvannet. Member D and Member E are first observed at the westernmost tip of Furuholmen (610750 20750). Both wedge out E of Fløosvann (611200 21500). Member F starts at the coast SE of Furuholmen. W of Terningvann (611000 21450) it gradually interfingers with Member K. For Members G and I there are gradual transitions into Member K south of Balsneshusvann. In the same area Members H and J wedge out.

The rest of the Vollan Formation is made up of Member K, which stretches all the way to Grindvik. It is possible that some of the sediments along the southern margin of the area belong to Member F, but there is no chance to determine this. Therefore they are all ascribed to Member K.

4.9.9 Member C

The thickness reaches a maximum of 240 m N of Furuholmen. Eastwards it thins and swells until it disappears W of Aksethusvann. The member is made up of sandstones, sometimes with conglomeratic horizons and channel fills. The diameter of the cobbles is seldom more than 10 cm, and they are always well rounded. No mudstone beds were detected within this member. The sandstones have a yellowish to greenish colour. Generally it is a coarse to very coarse grained sandstone, but there may be transitions into medium grained sandstone. The composition is described in Ch. 4.9.5. There is little change throughout the member, though there is a slight increase in the number of parallel laminated beds upwards and eastwards (Ch. 6.3.1).

4.9.10 Member D

Member D comprises a sequence about 14 m thick at the western part of Furuholmen. Eastwards it wedges out. The member is composed of mudstones, siltstones and very fine to coarse grained sandstones. There is a general coarsening upwards tendency throughout. The composition of sandstones and mudstones is described in Ch. 4.9.5. Member D is the only place where mudstones occurs between Members B and F. This is the reason why it is classified as a separate member.

4.9.11 Member E

A maximum thickness of about 40 m was measured at the coast southeast of Furuholmen. Eastwards the member wedges out. It is mainly composed of coarse to very coarse grained sandstones. The lithology and colour is the same as for Member C, but there are fewer conglomeratic horizons and channel fills. The sandstones form a monotonous succession with no obvious breaks or vertical changes.

4.9.12 Member F

An accurate thickness at the coast measured from logs gives 157 m. Eastwards there are minor variations. The member displays a wide variety of sandstone and mudstone beds in rapid alternation. The only trend throughout is a slight tendency for sandstone beds to become thicker upwards along with a decrease in thickness of mudstone beds. There are many upwards fining and coarsening sequences with thicknesses from one decimeter to about 16 m (Ch. 6.3.3). A wide variety of sedimentary structures are present. Grain size ranges from homogeneous mudstone to very coarse grained sandstone with conglomeratic horizons. Lithology of the different size fractions is described in Ch. 4.9.5.

4.9.13 Member G

The thickness measured from logs gives 79 m. Eastwards it interfingers with Member K. Conglomerates, very coarse and coarse grained sandstones dominate the member. The sediments are relatively massive. In the middle there are some beds of parallel laminated, medium to fine grained sandstones with occasional silty laminae and thin mudstone beds. Conglomerate cobbles and boulders are well rounded. They may in rare cases reach a diameter of 1 m. In the conglomeratic sandstones cobbles are not larger than 10 cm. The conglomerates are matrix supported, but near the bottom of the member there are clast supported horizons. A typical clast composition is shown in Table 4.4. The composition of the other fractions is given in Ch. 4.9.5.

4.9.14 Member H

The member is 12 m thick, but wedges out eastwards. It displays a fairly monotonous deposit of dark coloured, sandy siltstones. However, there are sandstone beds and channel deposits up to 50 cm thick, which range in grain size from very fine grained sandstones to coarse grained sandstones. The member is otherwise characterized by an unusual large number of big calcareous concretions, the largest one measuring 100x15 cm. Lithology is as described in Ch. 4.9.5.

4.9.15 Member I

The thickness measured from logs is 127 m. The member is dominantly composed of conglomerates and structureless, conglomeratic, very coarse grained sandstones. In the lower part there are some siltstone beds, of which the thickest is about 125 cm. The conglomerates are matrix supported with well rounded cobbles and boulders. Petrography is described in Ch. 4.9.5.

4.9.16 Member J

Maximum thickness measured at the coast is 48 m. Eastwards it thins and disappears very rapidly. The member is dominated by dark coloured siltstones and sandy siltstones with occasional intercalations of up to 0.5 m thick, medium to coarse grained sandstone beds and channels. They occur especially near the top and the bottom. The lower part of the member is a fining upwards sequence. When approaching the top it becomes coarser again. Sandstone beds usually have an erosional lower surface, either sharp or undulating.

4.9.17 Member K

A large number of strike parallel faults made it difficult to measure the accurate thickness. NW of Balsnes it probably reaches a maximum of about 250 m. Going eastwards, most of the Vollan Formation is made up of Member K, and finally, in the eastern part of the area, all of it (Ch. 4.9.8). In this way the member in some areas is more than 600 m thick. The reason for the thickening is either that the members below wedge out or their characteristics disappear so that they cannot be differentiated from Member K.

The member is composed of medium to very coarse grained sandstones and conglomeratic sandstones. Going eastwards and upwards towards the Balsnes Formation there is an increase in the number of conglomeratic horizons. They do normally not reach thicknesses of more than 2-3 m, and no attempt was made to map them out. An exception is a large area with conglomerates 1 km to the SW of Terningvannet (614200 31000). The conglomerates wedge out rapidly. Cobbles and boulders are always well rounded. Composition of the different lithologies is given in Ch. 4.9.5.

Mudstones occur with thicknesses of up to 10 m. Some of them can be traced over distances of more than 1200 m. In the western part of the area many mudstones wedge out over short distances or suddenly disappear because of channel incision (Ch. 6.3.7).

4.10 BALSNES FORMATION (MEMBER L)

4.10.1 General

The formation was first described by Peacock (1965) as the Balsnes Conglomerate (Ch. 4.1), and constitutes the uppermost deposits of the Hitra Group. It is analogous to Siedlecka and Siedleckis' (1972) Member E. No further subdivision was possible.

4.10.2 Name

After the farm Balsnes at the westernmost exposure of the formation.

4.10.3 Type Area

The steep hills just to the east of the farm Balsnes (610700 23200).

4.10.4 Thickness

An estimated minimum thickness is about 200 m.

4.10.5 Lithology and Petrography

The Balsnes Formation is generally a very massive, grain-supported, polymict conglomerate. Table 4.3 shows maximum boulder size and mean maximum boulder size measured at seven different localities along the area. No clear trend is obvious, and there are significant differences from place to place. The largest boulder observed was found just to the east of Balsnes Farm, and measured 2 m across.

In some areas scattered lenses and layers of very coarse conglomeratic sandstone made it possible to measure strike and dip. Sandstone layers could be a few metres thick, the lenses only a couple of decimetres.

Pebbles, cobbles and boulders are rounded to well rounded, ellipsoidal or spherical, but show no preferred orientation. The clast composition of the conglomerates was surveyed at five different localities along the length of the area. The results are given in Table 4.4. No obvious trends are detectable. In addition, the composition of the matrix was checked in a sample from the NE of Langnes (Table 4.2, spec. nr. 2064). As can be seen, it is similar in composition to sandstones of the Vollan Formation. Also the textural descriptions in Ch. 4.9.5 apply.

Locality	Grid reference	Maximum (cm)	Mean Max. (cm)
Hestvika	619500 40250	100	69.5
Badstuvik	617150 37400	26	21.9
Moldhaugen	616250 35500	130	60.0
Melkstaden	614850 33800	50	28.1
Olsvik	612100 26250	82	46.0
Aksethusberg	613650 29750	80	50.4
Balsneshusberg	610850 23150	100	58.3

Table 4.3 Maximum and Mean Maximum boulder size for the Balsnes Formation measured at seven different localities.
The Mean Maximum has been obtained by taking the average diameter of the ten largest boulders in a small area.

Locality	Grindvik	Grindvik	Badstuvik	Aksethusberg	Bekkviktjern	Member G
Grid reference	619500/ 40250	619500/ 40000	617150/ 37400	613650/ 23150	611500/ 25250	610850/ 21650
Rock Type	%	%	%	%	%	%
Granite	33.0	24.5	26.0	49.0	9.5	52.0
Pegm. Granite				3.5		3.0
Diorite	18.5	11.0	11.0	4.5	20.5	1.0
Chert	1.5	2.5	2.5	3.5	1.0	
Jasper	1.0	3.0	4.0	0.5	3.0	
Gabbro	15.5	9.0	7.5	13.5	6.5	8.5
Basalt	6.0	20.0	13.5	7.5	15.5	3.5
Diabas	6.0	11.0	8.5	4.0	4.0	5.5
Andesite	1.5	5.0	1.5	0.5	3.5	4.5
Gneiss	2.0	1.0	1.5	10.5	11.5	4.5
Schist	3.5	2.0	0.5	1.0	2.0	1.0
Metasandstone	3.0		1.0	0.5	0.5	7.5
Pegm. Qquartz		0.5	12.0	0.5	4.0	
Granodiorite	2.5	3.5	4.0		15.5	1.5
Pyroxenite	2.5	1.0	0.5		1.0	5.5
Acid Tuff		0.5				
Rhyolite	0.5	3.0	1.0		0.5	1.0
Trondhjemite	2.0	2.5			0.5	1.0
Limestone			1.5	1.0	1.0	
Amphibolithe			0.5			

Table 4.4 Clast composition of conglomerates of the Balsnes Formation and Member G. Percentages have been obtained by counting 200 points at 5 cm intervals within a random area.

4.10.6 Boundaries

The lower boundary is either erosional or gradational (Chs. 4.9.7 and 4.9.17). To the south of Litlvågen the formation apparantly lies directly on basement, but the contact could not be seen. At Balsnes the westernmost boundary is an eastwards dipping, normal fault. There is no upper boundary to other rocks preserved.

4.10.7 Distribution

The Balsnes Formation is preserved in two large areas, one stretching from Balsnes to Akset and another from the south of Litlvann to Grindvik. In addition it lies in a small area NW of Kallarnes.

4.11 LITHOLOGICAL COMPARISON BETWEEN THE MEMBERS

A lot of detailed petrographical work has earlier been carried out by Peacock (1965) and Siedlecka and Siedlecki (1972). Therefore this will only be a summary of the differences between the members.

Member A is usually easily recognized because of its relatively monomict, local basement composition. Member B is similarly outstanding in appearance, and is characterized by its red to grey colour. When examined under the microscope it can be seen to be rich in detrital biotite and calcite cement. Biotite may reach 10-15 %, calcite cement 15-20% (Table 4.2, spec. nr. 18703). Member B has a composition not very different from basement rocks in the area, from which it is supposed to be derived (Ch. 6.2.2).

Sandstones of the different members of the Vollan Formation do not differ much. No important lithological differences can be pointed out. Instead grain size and member thickness is used to distinguish them. In addition, the presence of various sedimentary structures may be characteristic. Typical compositions of the Vollan Formation sandstones are given in Table 4.2. Fig. 4.8 shows a plot of sandstone compositions from the samples of

Table 4.2. The sandstone sample from Member 8 clearly differs from sandstones of the Vollan Formation and the Balsnes Formation in containing less feldspar, gneiss and granite. The Balsnes Formation differs from the conglomerates of the Vollan Formation by its much higher content of jasper, chert and limestone, and by containing smaller amounts of granite. In addition it is usually more massive.

C H A P T E R 5

SEDIMENTARY FACIES

5 SEDIMENTARY FACIES

5.1 INTRODUCTION

To figure out the environment of deposition, it was considered necessary to divide the sediments into different facies. This was done on the basis of sedimentary structures, composition and grain size. Often there are gradual transitions between them, and many facies have a range of internal variability. Each facies tells which processes operated during deposition, but not much about the depositional environment. Each facies is presented as a descriptive part and an interpretation. The interpretation is meant only to suggest different environments, in which the particular facies may occur, and is primarily based on hydrodynamic properties. To assess the environment of deposition, different facies associations were studied (Ch. 6). A total of 11 facies were recognized.

5.2 FACIES

5.2.1 1. Siltstone and Sandy Siltstone

Description Grain size ranges from siltstone to very fine grained sandstone with high silt content. Usually the sediment is massive, but often a weak, millimetre-scale lamination is visible, being the result of slight grain size variations. The laminae are fining upwards. Other sedimentary structures were not found, but sometimes the sediments have been affected by soft sediment deformation, either within well defined horizons or throughout. The latter is most common. When beds of this facies become thinner than 5 cm, they are grouped together with Facies 2. This is an arbitrary limit picked for convenience during logging. Beds of Facies 1 are generally not thicker than 20 cm, but may reach about 1 m. The average is approximately 10 cm. In the Vollan Formation the colour is dark grey, in Member B it is red and sometimes dark green. Often it contains carbonate concretions, especially Member B is rich in those. Facies 1 of Member B may also be bioturbated.

Interpretation Absence of traction formed current structures suggests that the sediment was deposited from suspension. The normally graded, horizontal laminae are results of higher fall velocities for the bigger particles during sudden incursions of sediment. Slack-water vertical accretion following overbank flooding from an adjacent channel is a probable mechanism (e.g. Collinson and Thompson, 1982; Leeder, 1982; Nemec, 1983). Deposition may also take place by the same mechanisms in abandoned channels and lakes.

5.2.2 2. Siltstone with Sandy Lamination, Ripples and Lenticles

Description This facies is distinguished by the occurrence of sandy laminae and lenticles of mainly fine grained sandstone in a sequence of silt and sandy siltstone. The sand laminae/layers are 0.1-5 cm thick. When thicker, they are classified as separate beds of Facies 3 or 4. Single sandstone ripples are 0.1-3 cm thick, on the average about 1 cm. The length may be several decimeters. Cosets are common. Mudstone beds thicker than 5 cm and without sandstone ripples are classified as separate beds of Facies 1. Thickness of Facies 2-beds varies from 5 cm to several metres depending on in which member they occur (Ch.6). There are gradual transitions into Facies 1 and 3 both upwards and downwards. The thin, sandy laminae deposited from suspension show normal grading.

The cross-laminated sand ripples mostly exhibit unidirectional current directions. They may be both single and connected, and often occur in discrete horizons. The horizons may die out laterally, or there may be transitions into rippled sand beds. Both thick lenses and flat lenses are found (Reineck and Singh, 1980). The flat lenses are characterized by a length/height ratio larger than 20, for thick lenses the ratio is less than 20. Sand beds sometimes show Type 2 climbing ripple lamination (Reineck and Singh, 1984; Collinson and Thompson, 1982), in which the angle of climb is less than the stoss side slope, giving erosion between sets. Most ripples are asymmetrical, but occasionally symmetrical forms occur. The background, in which the sandy laminae and lenticles are found, is similar to Facies 1 both with respect to grain size, structures and colour. Sometimes siltstone appears to drape the ripples. Facies 2 of Member B is often bioturbated. Calcareous concretions are commonly found in distinct horizons. Soft sediment deformation structures are common. Examples of Facies 2 are shown in Figs. 5.1 and 6.16.

Interpretation The facies is not diagnostic for one special depositional environment. It has been found in environments such as tidal flats (Elliott, 1978), on marine shelves (Johnson, 1978), in deltaic areas (Elliott, 1978 and Miall, 1984) and on floodplains (Collinson, 1978a and Walker and Cant, 1984). Most probably there was a continental setting with deposition on floodplains and in lakes (Ch. 6). The rate of sediment input fluctuated through time. Beds with climbing ripples indicate high input, while single ripples indicate depleted sediment supply. Bidirectional, symmetrical ripples are indicative of wave activity. They are always wave modified current ripples. During periods of restricted sediment supply or low energy mudstones were deposited from suspension.

5.2.3 3. Ripple Laminated Sandstone

Description This facies is completely dominated by ripple laminated, fine grained sandstone, sometimes with silty drapes (flasers). Thin mud drapes may also be present between individual laminae in cross-sets. An example of the facies is shown in Fig. 5.2. The ripples usually exhibit unidirectional current directions, and there are gradual transitions into Type 2 climbing ripples (Reineck and Singh, 1980). The facies grades into Facies 2 and/or Facies 4. On the surfaces of Facies 3 beds the different ripple morphologies can be seen (Fig. 5.3). Crestlines vary in form from sinuos out of phase to linguoid and cuspate (Collinson and Thompson, 1982). Some surfaces contain symmetrical ripples which show modification by bidirectional currents indicative of wave action. The single ripples are 0.5-3 cm high, up to 15 cm long, and form cosets 5-20 cm thick. Beds/cosets thicker than 20 cm were not observed. On the average they are 10 cm. In sections perpendicular to the current direction the ripples show festoon cross-bedding.

Interpretation The ripple laminated facies is a product of migrating lunate and lingouid ripples in the lower part of the lower flow regime. It corresponds to Reineck and Singh's (1980) "Small Ripple Bedding". The sand is moved by bedload traction. With increasing sediment supply there are transitions into climbing ripple lamination. Flaser bedding is representative of deposition during periods of quieter water. The flasers are not indicative of a specific sedimentary environment. The symmetrical ripples with bidirectional current components are probably wave modified current ripples. A facies of this kind is abundantly developed on tidal flats, shoals, upper point bars, levees, in lacustrine sediments, deep-sea sediments and fluvio-glacial sediments (Reineck and Singh, 1980). However, when taking into account the existence of climbing ripples and wave modified current ripples one is left with upper point bars, levees and lacustrine sediments (Table 5 in Reineck and Singh, 1980). For further discussion, see Chapter 6.

5.2.4 4. Parallel Laminated Sandstone

Description Parallel laminated sandstone consists of horizontally laminated, very fine to medium grained sandstone. The laminae are normally 1-2 mm thick, but may reach 1 cm (Fig. 5.4). Each lamina is defined by its grain size and heavy mineral content, and shows normal grading. Beds range from 5-100 cm in thickness, averaging 15 cm. Parting lineation was not observed by the present author, probably because of lack of good exposures on bedding plane surfaces. Vertically there may be gradual transitions into Facies 3, ripple laminated sandstone (Fig. 5.2). Facies 3 usually occur on top of Facies 4. Basal surfaces vary from undulating erosive to planar erosive or non-erosive.

Commonly there are not more than single sets of parallel laminated sandstone, but cosets with very low angle erosion surfaces between sets were observed at two places (Ch. 6). These examples clearly have a close relationship to and a gradation into parallel laminated sandstone. The angle of lamination dip is low, only one to five degrees. Erosion surfaces are flat to slightly irregular (relief up to 6 cm), and are draped with silty material. Parallel laminated sandstones above may contain small mudclasts. The sets show normal grading and are 10-30 cm thick. It is probable that also some of the thicker beds of parallel laminated sandstone are cosets. Erosion surfaces between sets were not observed, possibly because they are parallel to the lamination.

Interpretation This kind of facies can be produced both in the upper and lower flow regimes (Collinson and Thompson, 1982). Lack of parting lineation does not point to lower flow regime conditions, as it may be the result of bad exposures. Peacock (1965) did observe parting lineation. Facies 4 may occur in many different environments. It has been described on storm dominated shelves (Reinech and Singh, 1972), in flysch sediments (Bouma, 1962) and in fluviatile environments it can be abundant on levees, point bars, sandflats and sand bars (Coleman, 1969). On levees and in overbank areas it is common to find parallel laminated sandstone deposited from sheet flows and crevasse splays. Ephemeral streams are often dominated by parallel laminated sandstone deposited from sheet or stream floods (Tunbridge, 1981). Inversely graded beds were not observed. Inverse grading of laminae results from swash and backswash on beaches or is found on intertidal flats where tidal currents operate. Therefore these origins are excluded for the parallel laminated sandstone beds (Reineck and Singh, 1980).

The presence of very low angle cross strata which pass transitionally into parallel laminated sandstone has also been noted by Smith (1970). He observed low amplitude sand waves forming in shallow depth in the lower or transitional flow regime. They produced nearly horizontal strata, which superficially resembled those of the upper flow regime plane bed during slowly increased flood stages on a floodplain. Nemec (1983) described similar deposits as the result of flood-stage accretion on sandflats in braided rivers. For further discussion, see Chapter 6.

5.2.5 5. Massive Bedded Sandstone

Description The deposits of this facies have a grain size varying from medium to very coarse sandstone, and often contain pebbles and mudclasts. In rare cases there are cobbles with a diameter of up to 15 cm. Mudclasts may be up to 1 m across, and show signs of plastic deformation. Cobbles, pebbles and mudclasts are restricted to the coarse and very coarse grained portions. The massive bedded sandstones are structureless, and are probably deposited within channels. Many conglomeratic channels cut down into Facies 5 (Fig. 5.5). Typically cobbles, pebbles and mudclasts are deposited near the bottoms. Beds of this facies are always thicker than 15 cm. Channel complexes with a thickness in excess of 10 m are not uncommon. They may be completely dominated by massive bedded sandstones. Single channels sometimes show a tendency to upwards coarsening. There are gradual transitions into conglomerate both vertically and laterally. Also gradual transitions into Facies 6 (Fig. 5.6) and Facies 9 are found.

Interpretation This facies is deposited as channel fill during rapid sedimentation. Erosion of earlier deposited sediments was widespread, as evidenced by the large number of mudclasts. Lack of internal structures indicates that sediments were dumped as a homogeneous mass and buried before significant bedload movement could take place (Collinson and Thompson, 1982). Normal grading reflects the deceleration of the sediment-laden current and consequent decline in competence, with coarse grains settling first. Massive bedded sandstone is common within river channels (Reineck and Singh, 1984; Rust and Koster, 1984).

5.2.6 6. Horizontal Bedded Sandstone

Description The facies consists of coarse to very coarse grained sandstones with scattered mudflakes up to 5 cm across and sometimes blocks of mudstone up to 0.5 m in diameter. The lamination is far from as well developed as in Facies 4. The laminae are normally 1-4 cm thick, and are mainly a result of varying grain size. A tendency to current (parting) lineation was observed, but no clear development. Basal surfaces are erosive, a nearly planar base is most common (Fig. 5.7). Bed thickness varies from 20 cm to more than 400 cm with an average of about 50 cm. Occasionally cosets were seen. Thin, up to 4 mm thick, laminae of silty, very fine grained sandstone separate them. Some of the thickest sets appear to be more than 4 m. It is probable that most of these are cosets with obliterated erosional surfaces in between. Occasionally thin horizons of Facies 3, ripple laminated sandstone, occur within such beds. They are supposed to have developed on the surfaces of single sets. Laterally they wedge out rapidly, probably as a result of erosion. No averge set thickness could be obtained, but it has to be less than the average bed thickness (about 50 cm). On top of Facies 6 there may be a sudden change to finer grained facies, either Facies 4, 3 or 2. Also common are gradual transitions to Facies 7 or 10 or sudden transitions to Facies 7.

Interpretation The coarseness of the sediment and occurrence of current lineation point to deposition in the upper flow regime. In addition the weaker differentiation between laminae compared with Facies 4 indicates upper flow regime conditions (Singh, 1972). Facies 6 represents the upper plane bed phase in the upper flow regime. Commonly horizontal bedded sandstone is found as channel fill above erosional scours, on top of sand bars and on sandflats (Nemec, 1983). Deposition from ephemeral streams (sheet floods and stream floods) is very common too (Tunbridge, 1981).

5.2.7 7. Trough Cross-bedded Sandstone

Description This is a widespread facies throughout the succession. Grain size varies from medium to very coarse grained sandstone, and there are scattered mudclasts and cobbles up to 15 cm in diameter. Bed thickness varies from 15 cm to more than 700 cm with an average of about 100 cm. Individual beds are made up of a number of poorly to well defined sets, producing cosets. Thickness of sets varies from 4 to 50 cm, but is normally not more than 15 cm. The average is about 10 cm. Width of troughs may be up to 2 m, and they can be several metres long. The thinnest sets exhibit the finest grain sizes.

Usually lower surfaces of beds are erosive, and have lag deposits of coarser material. Erosion surfaces are concave upwards, but may also be flat. The laminae are also concave upwards, and foresets are enriched in mica (Fig. 5.8). Some troughs contain up to 5 cm thick, irregular, laterally discontinuous cappings of ripple cross-laminated sandstone (Facies 3), and Facies 7 may grade into Facies 6, 8 and 10.

Interpretation This type of facies is a product of the upper part of the lower flow regime. Sediments are usually laid down in channels with erosive bases. Bedforms of this kind are the result of migrating megaripples (Reineck and Singh, 1980), also called dunes (Collinson and Thompson, 1982). The currents scour trough-like depressions, which are successively filled in by sandstone. Trough cross-bedded sandstone is common in most environments with relatively strong currents, e.g. tidal channels, river channels and crevasse splays.

5.2.8 8. Tabular Cross-bedded Sandstone

Description In the same way as Facies 7, this facies consists of medium to very coarse grained sandstone. The coarsest beds usually contain scattered mudclasts and pebbles up to 1 cm in diameter. Beds may only have one single set (solitary set) or be cosets. Set-thicknesses vary from 8 to 115 cm, but are seldom more than 30 cm. On the average they are 20 cm. Cosets contain from two to four sets and are 15 to 115 cm thick with an average of 30 cm. Single sets can be traced laterally for more than 35 m, while other may wedge out rather quickly, e.g. over distances of 5 m. The cross-bedded units are mostly tabular, but may also occasionally be wedge-shaped. Lower erosion surfaces of cosets are flat and sub-horizontal. Erosion surfaces between sets are either flat or concave upwards.

The foreset laminae are either tangential or sigmoidal or a transition between the two. The foreset lamination is usually well defined by alternation of finer (thinner) and coarser (thicker) laminae, or by grading within the laminae. Maximum foreset dip is 25-30 degrees, and is also commonly of this order. Lack of three-dimensional exposures usually made it difficult to say whether beds belonged to Facies 7 or 8. However, there are gradual transitions between them. Figs. 5.7 and 6.15 show examples of Facies 8.

Interpretation This kind of bedding is developed during migration of straight-crested megaripples (sand waves) in upper part of lower flow regime (Reineck and Singh, 1980, Collinson, 1970, Smith, 1972). It is very common as various sandbars in channels of braided rivers (Miall, 1977). Nemec (1983) distinguished large, solitary, Gilbertian microdelta-type transverse bars filling depressions, and smaller, multi-storey units (cosets) formed due to aggradation on sandflats. Sandwaves have also been observed on point bars, beaches, shoals and in tidal environments (Reineck and Singh,1984). Development of bottomsets depends on the amount of sediment deposited from suspension. The origin of coarse and fine layers on foresets is explained by Smith (1972) as the result of smaller bedforms on top of the megaripples migrating over the crests and avalanching down the foresets.

5.2.9 9. Intraformational Conglomerate

Description The intraformational conglomerate facies consists of medium to very coarse grained sandstone with mudflakes and mudclasts scattered throughout. The sandstone may be conglomeratic, and there are gradual transitions into Facies 5, massive bedded sandstone or Facies 10, conglomerate. There are gradual transitions between these three. The most common intraformational conglomerate has mudclasts up to 20 cm in length. On bedding surfaces mudchips appear to be flat-lying. The beds usually wedge out over distances of 20-30 m, but may reappear laterally. Mudclasts show evidence of plastic deformation. These beds are 5-25 cm thick with an average of 10 cm. In one channel a much coarser intraformational conglomerate was found, with plastically folded and deformed mudblocks up to 1x7 m (Ch. 6.3.7). Eroded calcite concretions with a size of 10x60 cm were also observed. Fig. 5.9 and 5.10 show the typical appearance of the facies.

Interpretation Beds of this facies occur near the bottom of channels or between beds deposited in upper part of lower flow regime or in upper flow regime. They are products of erosion of muds and silts which were cohesive, but not lithified. The eroding currents were strong. Intraformational conglomerates are common in many environments with strong eroding currents. The 5-25 cm thick mudchips conglomerates are probably products of erosion by crevasses, while thicker, mud-block conglomerates are deposited as channel lag in deeper channels (Ch. 6).

5.2.10 10. Conglomerate

Description The conglomerates are either grain-supported or matrix-supported, and the matrix is always a coarse to very coarse grained sandstone. Boulders may, in extreme cases, reach 2 m in diameter in the grain supported conglomerates, in which no sedimentary structures have been found. The matrix-supported conglomerates rarely have boulders larger than 30 cm across. Maximum and mean maximum boulder size of conglomerates in the Balsnes Formation are given in Table 4.3. Descriptions of grain sizes of conglomerates of the different members are given in Ch. 6. Occasionally there are lenses and layers of very coarse grained sandstone within the conglomerates. Both trough cross-bedding and horizontal bedding are present within these. Only at two localities was imbrication observed (Fig. 5.11). The conglomerates are polymict, and cobbles and boulders are always rounded to well rounded (Figs. 5.11 and 9.6). Conglomeratic units are always erosively based.

The reason why only one conglomerate facies is defined, is that all kinds of transitions exist between the different types. There are gradual transitions into Facies 5, massive bedded sandstone. Beds were found that resemble debris flows, but they are vaguely defined, and no separate facies was generated.

Interpretation Thin, grain-supported conglomerate lenses are common at the bottom of channels, where winnowing by currents has removed the fines and left a lag deposit. Thicker conglomerate units in continental settings most commonly represent deposition on alluvial fans, mainly in braided river channels. The high degree of rounding indicates transport and abrasion by river currents. Winnowing of fine grained material was widespread, as matrix is never finer than medium grained sandstone. This, combined with the almost total absence of sedimentary structures, point to a subaerial fan environment (Wescott and Etheridge, 1980). Finer conglomerates containing sand lenses and sand layers probably represent more distal parts of fans, where there was deposition of sediments eroded/transported from the proximal areas. Braided streams were probably responsible for these deposits too (Rust and Koster, 1984).

5.2.11 11. Sedimentary Breccia

Description The sedimentary breccia is characterized by angular, subangular and subrounded fragments ranging from a maximum of about 2 m in diameter to sand sized particles. The sediment is very poorly sorted. No sedimentary structures were observed except for a tendency to normal grading throughout the deposit. The largest fragments are found near the bottom, and finer, more sand-supported breccia occur near the top. There is a vague tendency for fragments to become better rounded upwards. On top there is a rapid transition into Facies 5 and 6. Facies 11 always rests unconformably on basement, often on a regolith (Ch. 3.3.2). Near the bottom of the deposit fragments are almost entirely composed of locally derived material, upwards more exotic fragments were found as well. Figs. 5.12 and 4.1 show examples of the facies.

Interpretation The coarseness and angularity of the clasts indicate short distance transport of sediments. Near the bottom the monomict composition shows that there was little influx of exotic material. Prior to deposition there were hills with an elevation in the order of 100-200 m (Ch. 6.2.1). Perhaps was the sedimentary breccia in some areas deposited as talus cones along slopes of these hills (Reineck and Singh, 1980), but generally the lower part of the member originated as an in situ weathering deposit. Upwards there is better rounding, perhaps as a result of weathering. However, the increasing polymict composition points to some influx of material by rivers. Generally the facies is described as an in situ basal weathering breccia, but the great thickness, in some areas up to 175 m, has to imply influx of sediments from nearby source areas. The transport does not need to have been long to obtain subrounded fragments.

CHAPTER 6

SEDIMENTARY ENVIRONMENTS

6 SEDIMENTARY ENVIRONMENTS

6.1 INTRODUCTION

During deposition of the Hitra Group a range of processes were active, of which many acted in different sedimentary environments. Generally each member represents a special environment or a set of environments. Therefore, for convenience, each member is discussed separately, or, if members display the same sedimentary environment, they are discussed together.

All logs, exept for those in Figs. 6.1 and 6.4 refer to the boundary between Members C and D as 0 m.

6.2 AUNE FORMATION

6.2.1 Member A

The formation starts with Member A, which is entirely made up of Facies 11, sedimentary breccia. It is the only member where this facies occurs, and therefore the facies description applies directly. The lower part of the deposit with relatively badly rounded fragments, and a monomict, local basement composition probably represents an *in situ* weathering breccia. Around elevated areas talus may have been deposited. However, as mentioned in the interpretation of the facies, the sometimes considerable thickness, and the existence of exotic fragments higher in the deposit must indicate influx of sediments. No sedimentary structures indicative of water transport were observed. This may be due to poor exposures. Only east of the mouth of Aunåa were exposures good (Ch. 4.8.6.1). Periodic floods and short-lived rivers along with small debris flows on small alluvial fans were the most likely transporting agents during deposition of the upper part of the member.

Topographic relief during deposition was in the order of 100-200 m. This has been established from the maps along the northern margin of the area (Plates 1A and 1B). When rotating the sedimentary pile back to horizontal, the undulating relief becomes clear. Orientation of bedding is sometimes sub-parallel to the unconformity. This is most pronounced close to the unconformity, and is therefore believed to be a result of later compactional effects (Peacock, 1965). Relief in the order of 100-200 m was enough for construction of alluvial cones. Member A was predominantly laid down in palaeovalleys (Plates 1A and 1B). Possibly there was some fault activity during deposition (Chs. 7.2 and 8.9.1), but this was not of major importance.

6.2.2 Member B

Above Member A there is a rapid transition into Member B. This member has also preferentially been laid down in topographical depressions, but covers a much wider area than the basal breccia. A representative log through part of the sequence was made from south of Storvågen (Fig. 6.1). Figs. 6.2 and 6.3 illustrate the measured section. The vertical organization of facies is rather complex. For purposes of simplicity a "Facies Relationship Diagram" (Harper, 1984; Selley, 1970) has been constructed (Fig. 6.40a). Typical upwards facies transitions are: Facies 9, intraformational conglomerate which passes gradually into Facies 7, trough cross-bedded sandstone or Facies 4, parallel laminated sandstone. Then there are gradual transitions into Facies 2, siltstone with sandy lamination, ripples and lenticles, Facies 3, ripple laminated sandstone and Facies 1, siltstone and sandy siltstone. Facies 3 often grades into Facies 7. There are rapid, gradual transitions between Facies 1, 2, 3 and 4, but few typical, repeated transitions. Usually there is an erosional surface on top of Facies 1, either planar or weakly undulating.

Very poor exposures made it impossible to trace beds laterally for more than a few metres. Exposures perpendicular to bedding are also poor, therefore it is not possible to say much about repeated occurrences of similar sequences. However, there are indications of small scale, 3-4 m thick, upward-fining "cycles". The lower surfaces are erosive, either planar or slightly undulating.

The lower portions are dominated by parallel laminated beds less than 1 m thick, often with rippled tops. Some beds contain a lot of trough cross-bedding. Between these beds there sometimes occur finer grained horizons of Facies 2.

The upper portions are dominated by Facies 2 and/or Facies 1. Finer grained sediments are often strongly bioturbated, and contain a large number of calcareous concretions, commonly arranged in distinct horizons. Desiccation cracks were not observed. It is common to find parallel laminated beds 1-2 m thick within the finer grained sediments. They have erosive basal surfaces with thin lags of intraformational mudclast conglomerate and often rippled tops. Beds of Facies 1 and 2 may reach thicknesses of 0.5 m and 2 m, respectively.

Smaller scale, 1-2 m thick, coarsening upwards sequences were also recorded. They may show upwards facies transitions from Facies 1, through Facies 2 to Facies 4 and/or 7.

No detailed study of palaeocurrent directions was carried out, but all ripples and cross-beds indicate palaeocurrents flowing in easterly directions.

Because of the few and poor sections available, an interpretation of Member B is not easily made. The 3-4 m thick, upward-fining "cycles" indicate waning strength of currents. They resemble the distributary channels with overbank sediments on top, which have been described by Pollard, Steel and Undersrud (1982) from the Devonian Hornelen Basin. Also Turner (1974) described similar looking cycles from the Late Silurian Ringerike Group in the Oslo Region as the product of meandering, distributary channels.

Lateral accretion surfaces do not exist in Member B, neither are there clearly outlined channel margins. The lack of steep channel margins, strongly erosive bases and the existence of Facies 2 beds between the coarser beds makes it unlikely that the "cycles" represent high sinuosity channels. Neither is it likely that they are low sinuosity channels, as there are few stacked, erosive based, coarse grained beds. The reason for the absence of larger channels may be lack of vegetation which could stabilize channel margins in Siluro-Devonian times.

The parallel laminated and trough cross-bedded beds probably do not represent crevasses, as crevasses originate from channels during breakage of levees, and channels were not observed. The common occurrence of single, erosive based, Facies 4-dominated beds is more probably an indication of sheet floods. In the distal reaches of ephemeral streams channels split up, and sediments are shed onto the floodplain during floods (Tunbridge, 1981). The strongest indication of this process is the predominance of parallel laminated sandstones. The parallel laminated beds in the tops of the "cycles" are also interpreted as sheet flood deposits (Fig. 6.1).

The finer grained sediments between beds of Facies 4 and 7 and in the tops of the "cycles" are interpreted as representing floodplain/floodbasin deposits. They are strongly bioturbated - an indication of rather slow deposition. The bioturbation introduces an age problem, as continental bioturbation should not be expected in the Silurian (Chs. 4.7 and 7.1). However, the organization of sedimentary facies does not indicate marine influence. Therefore the sediments are either Devonian, or continental bioturbation did occur in Late Silurian times. Micro-fabrics of the calcite nodules were examined by Siedlecka (1977), and were interpreted as being representative of freshwater precipitation. She interpreted the calcite nodules to be incipient palaeosols formed in an arid/semiarid climate. This seems to be a reasonable assumption, but it is still a problem why desiccation cracks did not form simultaneously, as palaeosols need a dry climate for their formation. The red colour of Member B is another indication of a continental, or shallow water, oxidizing environment.

The author's opinion is that the tendency for cyclicity within Member B is a result of switching of the ephemeral streams' distributary channels.

Upward-fining "cycles" were produced during shifting away from the particular area, while upward-coarsening sequences relate to shifting back of distributaries. From the east-southeasterly directed palaeocurrents it is probable that alluvial fans were situated to the N, W and S, from which ephemeral streams shed sediments into a floodbasin situated farther to the E, in the Hitra Area.

6.3 VOLLAN FORMATION

6.3.1 Members C and E

Members C and E will be discussed together because they are very similar both with respect to lithology, sedimentary structures and suggested depositional environment. What is stated for one member is also valid for the other.

The vertical organization of facies is illustrated in the "Facies Relationship Diagram" (Fig. 6.40b). Typical upwards facies transitions are: Facies 10, conglomerate grading into Facies 5, massive bedded sandstone, Facies 7, trough cross-bedded sandstone or Facies 6, horizontal bedded sandstone. Occasionally Facies 8, tabular cross-bedded sandstone was found above Facies 10. Facies 5, 6 and 7 occur in all positions with respect to one another. The lower surface of Facies 10 is always an erosional scour, the top is normally graded. Lower surfaces of Facies 7 are either erosional or gradational. All other surfaces are usually gradational.

A short log and a section were made from Member B at Aspvikodden, where exposures are good (Figs. 6.4 and 6.5). Fig. 6.6 is a photograph from the same area. Palaeocurrent directions were measured from trough cross-bedding on bedding surfaces (Fig. 6.7) at two localities. At Aspvikodden (Member C) there was a spread from 20 to 140 degrees, which gave a vector mean of 60 degrees (Fig. 6.8). On Furuholmen (Member E) there was a spread from 60 to 160 degrees with a vector mean of 110 degrees (Fig. 6.9). Members C and E probably represent river deposits. Many indications of channels were found. They are never more than two metres deep, and usually have a 5-10 cm thick, conglomeratic lag at the bottom. In addition there are channel conglomerates up to 1 m thick which are laterally persistent for at least 20-30 m. Exposures did not allow them to be traced over longer distances. The vertical arrangement of facies, the relatively little spread of current directions, only 100-110 degrees, and the shallow channels all indicate a braided river deposit.

Trough cross-bedded units dominate the deposits. They are here interpreted as representing migrating dunes in the deeper parts of channels. During scouring and aggradation channels become filled with this kind of bedform. Dunes have also been observed on top of tabular cross-bedded sets. This transition was interpreted by Nemec (1983) as the result of passage of sinuos crested dunes over a "unit" sand bar, a chute bar or an infilled channel floor depression during the passage from lower to upper flow regime (Leeder, 1982).

Sandwave deposits are present as tabular cross-bedded units not more than 25 cm thick. They are not common, but represented throughout. Usually there are only solitary sets occurring within thicker cosets of trough cross-bedded sandstone. The restricted thickness and occurrence as solitary sets within Facies 7 indicates that these sandwaves formed as minor bars within shallow channels or on sandflats during flood falling stages (Nemec, 1983).

Larger transverse and oblique bars which migrate within larger channels (Collinson and Thompson, 1982; Smith, 1974; Rust and Koster, 1984) were not observed. Similarly, sandflat aggradation was probably of minor importance, as such deposition is commonly dominated by multi-storey units of tabular cross-bedded sandstone (Nemec, 1983).

Horizontal bedded sandstones are interpreted as having aggraded on channel floors during flood stage (Nemec, 1983).

There is a complete lack of mudstone beds in Members C and E, but there are mudstone clasts and silty drapes on the foresets of cross-beds. It is likely that mudstone layers and beds were deposited in small ponds and abandoned channels in the river system, but they had a low preservation potential because of frequent channel migration/shifting.

When Member C is traced eastwards to the area south of Grunnvann, there is an increasing abundance of Facies 6, horizontal bedded sandstone, which may pass laterally into Facies 4, parallel laminated sandstone. A different depositional process probably dominated this area. A possible explanation is that there was a higher frequency of longitudinal bars, on which these two facies are common (Walker and Cant, 1984).

Facies 6 was perhaps also deposited as sheet flood sandstones. According to Tunbridge (1984), the abundance of horizontal bedded sandstone is consistent with an ephemeral stream dominated by sheet flood and stream flood processes. The lack of fine grained sediments point to the proximal part of an ephemeral stream with rapidly migrating, low sinuosity channel systems sweeping the alluvial plain. Repeated episodes of channel cutting, vertical accretion, avulsion and sheet flood deposition could account for the sequence that is seen. Where trough cross-bedding dominates the sequence, more constantly flowing, possibly perennial streams were active (Tunbridge, 1981).

To conclude, Members C and E were deposited in the middle portion of an alluvial fan, mainly by braided rivers. Eastwards ephemeral stream deposition probably dominated during sedimentation of Member C.

6.3.2 Member D

Member D is discussed in a separate section because it is the only mudstone sequence occurring between Members B and F. The typical facies transitions are illustrated in Fig. 6.40c. The member starts with Facies 1, siltstone and sandy siltstone which passes gradually into Facies 2, siltstone with sandy lamination, ripples and lenticles. Then there is an increasing abundance of Facies 3, ripple laminated sandstone and Facies 4, parallel laminated sandstone within Facies 2, before ending up with Facies 7, trough cross-bedded sandstone above a planar erosional surface (Fig. 6.10). There are no erosive channel margins. A 1 m thick bed of Facies 2 caps the member.

In the middle of the sequence there are many 20-50 cm thick, fining upward sequences with basal, planar, erosive surfaces. They start with beds of Facies 4, which may be up to 15 cm thick. Then come 1-10 cm thick beds of Facies 3 and finally Facies 2. Facies 3 may also occur as up to 10 cm thick, isolated beds within Facies 2. Climbing ripples are common within these. Sometimes the ripples have silty drapes which pass into flaser bedding.

The general picture of Member D is a coarsening upwards sequence. This implies a shallowing depositional environment. Considering the members above and below, which represent braided rivers (Ch. 6.3.1), the most likely environment would be a lake on the braidplain related to switching of the main distributary. This led to deposition of only fine grained sediments. As the main channel system returned back, an increasing amount of coarser grained material was deposited in the lake, mainly during flood events.

Instead of the lake model there is the possibility that Member D was deposited in an abandoned channel. Since no channels deeper than 2 m have been observed in Members C and E (Ch. 6.3.1), the author tends to prefer the lake model.

The small fining upwards sequences in the middle of the member is probably related to breakage of channel margins during floods. Sediment-laden water was shed across the surface onto the braidplain/lake area as crevasses. When entering the lake they behaved much as low density turbidity currents, and the typical 20-50 cm thick upwards fining facies transitions resulted (Bouma, 1962).

The sudden incoming of a 1 m thick bed of parallel laminated sandstone near the top is probably the result of one single flood event. This implies that the lake was already filled up with sediments. However, it was probably never very deep. It is also possible that the parallel laminated bed simply represents a vertically accreted sandstone at the bottom of a channel deposited in the upper plane bed phase during upper flow regime conditions. The lack of scour and mudclasts near the bottom favours the flood model. On top of the parallel laminated sandstone there is another sudden transition, now into trough cross-bedded channel sand.

An alternative to the established model for deposition of Member D is the possibility that a small delta periodically supplied sediments into the lake, filling it up. The parallel laminated sandstone bed near the top could then either represent a small mouth bar or a sandbar of a low sinuosity river which migrated across a low energy shoreline (Pollard, Steel and Undersrud, 1982).

6.3.3 Member F

6.3.3.1 Introduction

Member F represents the most complex depositional environment in the Hitra Group succession. There is a whole suite of different facies associations, which probably represent changing conditions during deposition. Siedlecka and Siedlecki (1972) reported a repeated occurrence of upward-fining cycles produced by meandering rivers on a floodplain. The cycles were recognized, but their interpretation is not considered correct, as will be shown. A total of 9 more or less distinct cycles were observed, but in at least two places there are coarsening upward tendencies that allow different interpretations. This will be discussed in Ch. 6.3.3.4. The 9 cycles mentioned have thicknesses in the order of 11-16 m.

In addition smaller coarsening and fining upwards sequences will be described and discussed.

6.3.3.2 The Sequence of Facies

A section throughout Member F is presented in Fig. 6.11. As can be seen, the lower surfaces of each cycle are not similar. Two types are discerned:

1. Undulating erosional overlain by interbedded Facies 5, massive bedded sandstone, Facies 10, conglomerate and Facies 9, intraformational

conglomerate. Any of these may lie directly above the contact. There is usually a gradual transition upwards into Facies 6, horizontal bedded sandstone.

2. Planar erosional overlain by Facies 7, trough cross-bedded sandstone, Facies 6, horizontal bedded sandstone or Facies 8, tabular cross-bedded sandstone.

More than one erosional surface may occur in the lower part of each cycle. Above an erosional surface the general facies trend is: Facies 5 (with horizons of Facies 9 and/or Facies 10). Then Facies 6, 7 or 8, which may occur in all positions with respect to one another. A "Facies Relationship Diagram" is presented in Fig. 6.40d.

Then there is a rapid transition into finer grained sediments of Facies 4, parallel laminated sandstone, Facies 3, ripple laminated sandstone, Facies 2, siltstone with sandy lamination, ripples and lenticles and Facies 1, siltstone and sandy siltstone. The last mentioned occur only sporadically. The most common arrangements are: Facies 4 above a planar or slightly undulating surface that may be erosional or not. Then comes Facies 3 and finally Facies 2. The transition from Facies 4 to Facies 3 sometimes goes via wavy bedding. Any or both of Facies 2 and 3 may be absent. Occasionally Facies 8 is found interbedded with Facies 4, and it is not uncommon to find beds of Facies 7 and Facies 9 in the upper parts of the cycles. Facies 2 and 4 dominate the upper parts of the cycles, Facies 6 and 7 are volumetrically most important in the lower parts.

Occurring within these large fining upwards cycles there are smaller sequences in the order of 50-350 cm. They have the same vertical arrangement of facies, but Facies 5, 6 and 10 are not developed. Lower surfaces are either erosional or non-erosional. The smallest fining upwards sequences are made up of Facies 4, 3, 2 and 1. They are 10-20 cm thick. Small coarsening upwards sequences in the order of 10-100 cm also exist within the 11-16 m thick cycles. They have the same facies as the 50-350 cm thick sequences mentioned above, but with an inverse arrangement.

6.3.3.3 Description/Interpretation

When trying to interpret the sedimentary environment it is important to consider the general setting, that is, to consider members above and below. Member E has been interpreted as the deposit of braided rivers in a mid-fan area, Member G (Ch. 6.3.4) is a similar deposit. The most likely model for Member F would then be a distal equivalent to a braided river, that is a floodbasin. In such an area, small lakes may have developed periodically with rivers feeding in sediments.

Current directions were measured from trough cross-bedding (Fig. 6.7) at three different localities. Only unambiguous measurements from rib and furrow structures on bedding surfaces were used, and few were obtained, but they all give a clear picture of currents flowing about 75-80 degrees (ENE). There is little spread of the measurements (Figs. 6.12, 6.13 and 6.14). A similar trend was observed from ripples and tabular cross-beds. The fairly constant current directions do not indicate that the 11-16 m thick cycles represent high sinuosity river channels, as one then should expect larger variation from place to place (Collinson, 1978b; Smith, 1972; Thompson, 1970). No typical point bar sequences were observed.

For the 11-16 m thick cycles the author is more tempted to believe in a model involving low sinuosity rivers in a floodplain/distal alluvial fan environment. The presence of multiple stacked erosion surfaces in the lower parts of the cycles may be due to scour and fill within a non-migrating channel, and need not necessarily indicate frequent channel migration (Collinson, 1978b). Some of the channels show scour, some do not. Migrating within the channels were dunes and sandwaves, now represented by trough cross-beds and tabular cross-beds, respectively (Figs. 5.7 and 5.8).

Member F is the only member where relatively large scale tabular cross-beds are found. Usually each set is not more than 30 cm thick, but one set 115 cm thick was observed in the lower part of Cycle 6 (Fig. 5.7). The sigmoidal foresets and relatively large thickness may indicate that it was deposited as a transverse or an oblique bar within a wide channel

(Nemec, 1983). It is also possible that it represents a small delta-like bar that grew at the mouth of a local chute (Miall, 1977). Laterally the thickness of the sandwave diminishes rapidly, and another set starts to grow. This may be the result of a period of stagnation before reactivation. Other possibilities are discussed by Nemec (1983).

In the lower part of Cycle 3 there is a stacked sequence of tabular cross-bedded sandwaves and parallel laminated sandstone beds which probably represents aggradation on a sandflat (Fig. 6.15). These are broad shoals in river channels, developed through the coalescing of sandbars onto one another (Cant and Walker, 1978). Aggradation mainly takes place during floods. Sandflats are well described by Nemec (1983) and Cant and Walker (1978).

Horizontal bedded sandstone is either a product of upper flow regime sedimentation within channels, on sandflats or deposition on longitudinal bars.

Collinson (1978b) mentions the problem of discriminating between low sinuosity streams and sheet flood/crevasse splay deposits. It is probable that the lower parts of some of the cycles, e.g. Cycles 4, 7 and 9 represent deposits of the last mentioned type. They all show flat, basal surfaces with little scour. Three metres above the base of Cycle 4 there is a layer of massive, black mudstone, which is eroded on top and wedges out laterally. This is the result of a period with deposition from suspension before subsequent erosion. Most likely the lower part of Cycle 4 is a stacked crevasse/sheet flood deposit exhibiting quieter periods between floods with deposition from suspension in a standing body of water. Whether or not the lower parts of other cycles represent stacked crevasses/sheet flood deposits or low sinuosity river channels is difficult to say, because poor lateral exposures makes it impossible to deduce anything about sand body geometry (Ch. 6.3.3.4). The problem of discriminating between sheet flood deposits and crevasses is further discussed in Ch. 6.3.3.4.

The 50-350 cm thick upward-fining sequences within the larger cycles all show erosional scours at the bottom, and contain intraformational mudclast conglomerates. Some appear to wedge out in an easterly direction. The upward-fining is evidence of waning flow. Similar deposits were described by Coleman (1969) as crevasse splays being deposited on levees during overbank flooding. An example is the upward fining sequence in the middle of Cycle 3. Upward-fining crevasse channel deposits are also a common product of overbank flooding (Coleman, 1969). An example is the upper part of Cycle 7, which is dominated by crevasse channels 0.5-1 m deep (Fig.6.11).

The upper parts of the 11-16 m thick upward-fining cycles bear evidence of strong and repeated traction currents. Facies 1 is rarely developed. Facies 2 is full of fining upwards sandy laminations, ripples and lenticles (Fig. 6.16). Ripples and lenticles occur in discrete horizons. This facies represents overbank sediments on a floodplain, partly deposited in shallow lakes from suspension. Periodical floods led to breakage of channel margins and 10-20 cm thick, upward-fining crevasse splays were shed into interchannel areas. They nearly always start with a basal, planar or slightly undulating erosional surface, and are followed by parallel laminated sandstone which often pass gradually upwards into ripple laminated sandstone via a thin, wavy laminated horizon. There are transitions from ripple lamination into climbing ripple lamination. Finally the 10-20 cm thick sequences may be capped by thin layers of Facies 2. The fining upwards is a result of waning strength of traction currents. The presence of climbing ripple lamination (Fig. 5.2) indicates strongly suspension laden currents (Collinson and Thompson, 1982).

Some upper surfaces of Facies 3, ripple laminated sandstone, are reworked by waves, thereby giving ripples a symmetrical geometry (Fig. 5.3). When seen in cross section, ripples immediately below the top are asymmetrical, and exhibit unidirectional current directions. The reworking must have taken place in shallow lakes on the floodplain.

Some of the 10-20 cm thick, normal graded sequences show a massive appearance in the lowermost 1-2 cm that could be the result of vertical aggradation from suspension. Then there is a rapid transition into Facies 4, which is a product of traction currents. Inverse grading in the lowermost, massive part was observed a couple of places. Similar deposits were described from lake environments by Steel and Aasheim (1978) as small turbidites.

In addition to the described upward-fining sequences, there are upward-coarsening sequences with thickness in the order of 10-100 cm. Each sequence is composed of a number of the 10-20 cm thick, upwards-fining flood sequences described above. The coarsening upwards sequences are also part of the floodplain deposits, and are probably a product of levee progradation (Reineck and Singh, 1984). There is an upwards increase in the amount of ripples, climbing ripples and parallel laminated sandstones in addition to an increase in bed thickness. This kind of levee deposits are also described from deltaic environments (Elliott, 1978). Typical examples are shown in Figs. 6.17 and 6.18.

Levees are characterized by rapid deposition of large volumes of sediments, and may therefore be unstable (Elliott, 1978; Reineck and Singh, 1984; Coleman, 1969). Soft sediment deformation should be expected, and is found. It is especially well developed below thick beds of rapidly deposited, coarser sandstones, e.g. in the top of Cycle 3 (Figs. 6.19 and 6.20). They are products of instability, either because of increased overburdon or sudden movements in the sedimentary basin. Such movements might be the result of fault activity. Both mechanisms may cause the heavier sand to sink as ball and pillows or the lighter mud to raise as flames. Fig. 6.21 show the overall picture in the top of Cycle 3, which is interpreted as levee deposits.

6.3.3.4 Discussion

In summary, Member F is interpreted as low sinuosity channels/sheet floods/crevasse channels overlain by floodplain/levee/lake sediments deposited from suspension and by traction currents. Cutting down into the latter are smaller crevasse channels. However, as mentioned in the introduction, there are other possible interpretations. Sediments very similar to those described here are found in interdistributary areas of fluvial dominated delta plains, which in this case would be lake delta plains, as no marine indicators were found.

The interdistributary areas tend to be dominated by a wide range of facies and sequences that represent infilling by a variety of flood-generated processes (Elliott, 1974). Overbank flooding is an important process involving sheet flow of sediment-laden water over the channel banks. Fine grained, laminated sediments are deposited over the entire area. Close to the channels levee progradation takes place with deposition of Facies 3, ripple laminated sandstone and Facies 4, parallel laminated sandstone. Quieter periods are characterized by Facies 1, siltstone and sandy siltstone and Facies 2, siltstone with sandy lamination, ripples and lenticles. Crevasse splays will deposit their sediment either on the levees, or, if they protrude into a lake, they may develop into a density current and deposit an erosive based lobe of sand 2 cm to 2 m thick (Arndorfer, 1973, Steel and Aasheim, 1978). What is described here is very similar to the sediments in the upper parts of some of the cycles, e.g. Cycles 3 and 5. The two environments can hardly be held apart, as they develop similar structures and sequences.

The transitions between Cycles 5 and 6 and Cycles 8 and 9 show coarsening upwards tendencies in addition to repeated scour and fill represented by either erosional surfaces or conglomeratic horizons. This can be due to varying strength of currents and varying character of transported sediments within a non-migrating channel. Another possibility is migration of bar heads over bar tails within river channels (Bluck, 1976). However, the relatively large thickness, 4-5 m, is suggestive of larger scale progradation, probably involving levee/crevasse channel complexes into a floodbasin. This was followed by abandonment of the channel complexes, which resulted in deposition of finer grained sediments. The process was described by Elliott (1978) and Steel and Aasheim (1978).

If deposition took place in an interdistributary area, the two coarsening upward sequences described above could represent minor mouth bar/crevasse channel couplets. As a minor mouth bar/crevasse channel progrades into a lake, proximal facies progressively overlie distal facies. Muds and silts deposited in the lake pass upwards into interbedded silts and sands with multidirectional trough cross-lamination which represent current and wave action on the mouth bar front (Elliott, 1974). Clear multidirectional troughs were not observed, but current ripples are sometimes modified by wave action (Fig. 5.3). The upper part of such sequences are frequently removed by erosion at the base of crevasse channels as progradation continues. If the interdistributary lake model is correct, such removal must have been extensive. The author tends more to believe in the first model involving levees and crevasses on a floodplain, because:

1. No well developed mouth bar sequences were observed.

2. Current directions from ripples, planar cross-beds and trough cross-beds are unidirectional, only some slight wave reworking has taken place.

3. Associated with an interdistributary area one should expect to find prominent channels from which crevasses and sheet flows could originate during breakage of levees. Such are not found.

The last argument can also be used against the low sinuosity river/floodplain model. As already mentioned, it is difficult to say if the lower parts of the cycles really represent low sinuosity channels or erosively based crevasses/sheet floods. Fig. 6.22 shows frequency of sand beds (very fine grained sandstone-conglomerate) plotted against bed thickness. There is a fairly unimodal distribution, which indicates that all sand beds have been deposited by the same processes, that is, sheet floods/crevasses (highest peak). However, the author tends to believe that the two minor peaks at about 105 cm and 205 cm may represent low sinuosity channels.

If there are no bigger low sinuosity channels present one should not discuss crevasse splays and crevasse channels at all, because these have to be derived from larger channels. Then the most likely candidates for deposition of the coarse grained parts of the member would be sheet floods/ephemeral streams in the distal parts of alluvial fans. During quieter periods fines would settle out. In such an environment one should not expect to find well defined, repeated fining upward cycles. Instead there should be an unsystematic alternation of thin and thick fining upwards sequences deposited mainly by verical aggradation. However, at least in the lower part of Member F there are well defined, fining upwards cycles representing major channeling events. Channels were filled in during several depositional episodes.

The predominance of trough cross-bedded sandstone in the lower parts of the cycles indicates rather persistent flow. Ephemeral streams/sheet floods are completely dominated by parallel laminated and horizontal bedded sandstone (Tunbridge, 1981). Another argument against the sheet flood model is that the mudstones between coarser beds contain large numbers of sand ripples and lenticles indicative of frequent current activity. In a sheet flood/distal ephemeral stream environment fines are usually massive. Only during flood events sandstone is brought into the basin. Between floods fines are deposited from suspension in ponds and temporary lakes (Tunbridge, 1981).

In conclusion, amongst the various models, the distal alluvial fan/low sinuosity river/floodplain model appears to be most acceptable.

6.3.4 Members G and I

Members G and I are discussed together because they are similar with respect to sedimentary structures and suggested depositional environment. Member I is coarser and more massive than Member G. A section through Member G is presented in Fig. 6.23.

Many more or less important erosional surfaces are present. Starting at the base, typical facies transitions between the main surfaces are (Fig. 6.40e): Facies 5, massive bedded sandstone, Facies 9, intraformational conglomerate and Facies 10, conglomerate passing into Facies 7, trough cross-bedded sandstone and Facies 5 with occasional horizons of Facies 6, horizontal bedded sandstone, Facies 9 and Facies 10. Any of these facies may be absent. Often Facies 6, Facies 4, parallel laminated sandstone and Facies 2, siltstone with sandy lamination, ripples and lenticles are preserved on top. Facies 5 and 10 constitute most of the members. Because of erosion only very thin beds of Facies 2 are preserved.

It is typical for beds to wedge out over relatively short distances, e.g. 20 m. This is especially the case for conglomerate beds which represent the lower part of upward-fining channel deposits. Conglomerates are always matrix supported, and contain boulders up to 1 m in diameter. Near the base of Member G imbrication was observed in the conglomerates, giving a current direction to the E (Fig. 5.11). Similarly, a southeasterly directed current direction was obtained from the sand trail behind an isolated cobble in the massive sandstone of Member I (Fig. 6.24). A flow separation eddy has developed, allowing deposition of finer grained particles at the lee side of the cobble.

When Facies 6 occurs within Facies 5 or 7, it is usually not well developed. It is vaguely defined, and disappears over short distances. However, its occurrence is quite commom. Debris flow-like deposits are found within the massive bedded sandstones of Facies 5. They may be up to 50 cm thick. The lowest few centimetres are inversly graded, and at the top there is normal grading. The vaguely defined horizontal bedding which is ascribed to Facies 6, is probably often a result of small debris flows, but the debris flows are so poorly defined that no separate facies was created.

In some places parallel laminated, finer grained sandstones of Facies 4 are found within thicker units of Facies 5. The laminae may be silty. Beds of Facies 4 commonly overlie massive sandstones, which again may overlie up to 10 cm thick beds of intraformational conglomerates. These small upward-fining sequences are up to 20 cm thick. They are thought to represent erosively based sheet flood deposits and crevasses.

Member I is mainly made up of conglomerates and massive sandstones, and can not be further subdivided. Within Member G a fining upwards cyclicity was found between major erosional surfaces. Three or four larger "cycles" could be defined, and they could again be divided into smaller sequences. Each sequence starts with conglomerates, intraformational conglomerates and massive sandstones above a channel floor. There is an upward transition into massive sandstones with trough cross-bedding, horizontal bedding, small debris flow-like deposits and sheet flood deposits. Finally, at the top, mudstones deposited from suspension with occasional ripples happen to be preserved. The sequences represent channels which have been eroded and then subsequently filled with sediments during flood events. Abandonment or longer quiet periods led to deposition of fines.

Members G and I were probably deposited as lobes on an alluvial fan which prograded onto floodplain sediments of Member F. The coarse and relatively structureless appearance points to the mid/proximal part of an alluvial fan. There is a slight coarsening upwards tendency in the top of Member F, which may reflect this progradation. However, the author is of the opinion that the break between Members F and G represents a longer time span. Progradation had not yet started during deposition of Member F (Ch. 4.9.7). Member G represents the sudden incoming of coarse river sediments and mass flow deposits. Deposition of these continued throughout the accumulation of Member I, only with a break during deposition of Member H.

6.3.5 Member H

This member is described separately even though it is very similar to the thin mudstones appearing within Members G and I. It is, however, much thicker. A section throughout the member is presented in Fig. 6.23. The absolutely dominating lithology of the member is Facies 2, siltstone with sandy lamination, ripples and lenticles. Cutting into this lithology are up to 0.5 m deep channels with massive, medium to coarse grained sandstones of Facies 5. In addition there are thinner, flat based sand sheets of the same lithology.

Small sand diapirs (Fig. 6.25) and soft sediment deformation structures are common. They result from sediment instability after rapid deposition. Deposition from suspension was most important for the finer grained lithologies, as ripples and lenticles are relatively scarce. Also graded lamination is common.

The member probably represents deposition on the more distal portion of an alluvial fan lobe (Ch. 6.3.4), possibly periodically a shallow lake. During floods sand was shed into the area, and sandstones were deposited as massive, flat based sheets. Stronger erosive currents developed channels, which were filled by massive sandstones.

Another possibility is that Member H was deposited in an abandoned channel. The sediments would be much the same. However, if the channel was plugged, it was perhaps filled with water, and crevasses could not erode channels down into the finer grained sediments. Crevasses would rather be deposited as flat based sand sheets and small turbidites. Therefore it is considered most likely that Member H is a floodplain/temporary lake deposit.

6.3.6 Member J

Member J constitutes the thickest fine grained sequence in the Hitra Group. As can be seen from Fig. 6.26, the lower part of it is a fining upwards sequence (Fig. 6.27) because of fewer and thinner sandbeds upwards. The upper part is a coarsening upwards sequence, as there is an increased number and thickness of sandbeds. Most of the member is made up of Facies 2, siltstone with sandy lamination, ripples and lenticles. Cutting into this are flat or undulating, erosively based sandstone sheets of Facies 5, massive bedded sandstone or Facies 6, horizontal bedded sandstone. One bed, near the bottom, showed Facies 7, trough cross-bedded sandstone. Near the top a 50 cm deep channel was observed. The coarser sandstone beds near the top or the bottom are normally graded, 10-50 cm thick and wedge out over distances of 20-30 m. They commonly merge together when traced laterally. Finer grained sediments in between disappear as a result of erosion. The general trend near the top is that sandstone beds merge westwards. Eastwards they split and disappear. Sometimes they show horizons of Facies 9, intraformational conglomerate near the base. The thicker beds near the top may contain mudclasts up to 0.5 m across.

On top of sandstone beds there is an abrubt change to Facies 2. The tops are usually rippled. Current ripples dominate, but there is evidence of wave modification seen by symmetrical ripples. Flaser-bedding was observed in horizons with symmetrical ripples. From the asymmetry of ripples seen at bedding surfaces, currents appeared to have been directed towards the

SE. However, some of the ripples show signs of tectonic modification, so they can not simply be used as current direction indicators (Fig. 6.28). Anyway, some of the ripples were measured and plotted in a sterogram (Fig. 6.29). After rotation back to horizontal, the ripple crests had a trend NE-SW, that is, currents were either directed to the NW or to the SE.

The general fining upwards trend near the bottom of the sequence and the coarsening upwards trend near the top is also reflected in the number and thickness of graded sandstone layers within Facies 2. In the middle of the sequence they are not more than 2 cm thick. Going up or down, the thickness may increase to 4-5 cm. Usually they are laterally persistent for at least 10-20 m, have rippled tops and are normal graded. In the middle of the sequence laminae may be extremely thin, suggesting deposition from suspension. Ripples and lenticles occur in discrete horizons. Soft sediment deformational structures are common in Facies 2, especially near the top and bottom of the member (Figs. 6.30, 6.31 and 6.32).

Member J probably represents an interdistributary lake deposit . The fining upwards sequence at the bottom is the result of gradual abandonment of sediment supply and increasing water depth, probably because of lobe switching. Sandstone beds represent crevasse splays. Crevasse splays which reached the standing body of water developed into small turbidity currents, which deposited the flat based, normally graded sandbeds with rippled tops. The lake was never very deep, as current ripples often are modified by wave action.

The top of the member bears evidence of shallowing conditions and increased sediment input. Crevasse splays are common. The coarsening upwards sequence does not resemble a distal mouth bar/delta front deposit as described by Elliott (1974). It is possible that the large channel at the bottom of Member K has eroded away the mouth bar, but it is considered more likely that the lake gradually filled and developed into a shallow lake flat.

Derivation directions for the sediments are uncertain for this member, but indicators for transport towards the SE are:

1. Asymmetric current ripples indicating palaeocurrents towards the SE.
2. Eastwards thinning of crevasse splay sandstones.
3. Syn-sedimentary, listric, normal faults dipping to the SE.

6.3.7 Member K

This member is made up of coarse to very coarse grained sandstones with thinner conglomerates and mudstones between the sandstone beds. Because of bad exposures no attempt was made to measure a section through it. The depositional environment varied during sediment accumulation. Most of the succession is made up of facies sequences which have been described earlier. The sedimentary environments were probably not very different, so references will be made to earlier member descriptions.

The dominating facies throughout are Facies 5, massive bedded sandstone and Facies 7, trough cross-bedded sandstone. The most common arrangement of facies in the coarser portions is matrix supported conglomerates and intraformational conglomerates above a basal, undulating, scoured surface. Conglomerate cobbles are rarely larger than 15 cm in diameter. Bimodal conglomerates have been observed (Fig. 5.5). The basal conglomerates vary in thickness from only one pebble thick to 3 m. Then follow Facies 5 and finally Facies 7. Very often Facies 7 is not present. These sequences represent braided river channel deposits (Chs. 6.3.1 and 6.3.4). The width of channels has at places been measured to be more than 30 m. Usually a number of the braided channels are stacked on each other, and bimodal conglomerates indicate that frequent redeposition took place.

The other common facies sequence within Member K has a lower portion, which, above a scoured channel surface, starts with Facies 10, conglomerate and Facies 9, intraformational conglomerate. Then follow Facies 5, massive bedded sandstone and occasionally Facies 7, trough cross-bedded sandstone. On top mudstone sequences mainly made up of Facies 2, siltstone with sandy lamination, ripples and lenticles are preserved. Within Facies 2 normally graded, sandy laminae dominate, and soft sediment deformation structures are common (Fig. 6.33). Cutting into this facies are crevasse channels and splays of the same type as those occurring within Member H (Ch. 6.3.5). The crevasses contain massive bedded or parallel laminated, fine to coarse grained sandstones, and are not more than 15-20 cm thick.

These upward-fining sequences represent channel deposits with overbank mudstones on top. Channel deposits with overbank fines between may be stacked on top of each other (Fig. 6.34). Whether the channels were high or low sinuosity is difficult to say. Some of them cut into overbank sediments with channel margins dipping as much as 45 degrees (Fig. 6.35). This could suggest that at least some of them were high sinuosity channels with strong erosion in the outer bends of the meanders. Typical point bars as described by McGowen and Garner (1970) are rarely observed. Lateral accretion surfaces are absent, and channel deposits are commonly massive. However, what is interpreted as a point bar sequence is shown in Fig. 6.36.

Some channels were obviously very deep. Channels cutting 10-15 m down into overbank deposits are not uncommon in the Balsnes area, but some of the deepest channels were observed to contain several erosive surfaces, and are probably stacked sequences of shallower channels. Figure 6.37 is a drawing of the lowest part of Member K at the coast W of Balsnes. The boundary between the two members is the bottom of a channel with large, plastically deformed mudstone blocks. The feature was first observed by Peacock (1965), who used the name "detached masses". He described the deposit as a result of water injection and subsequent flowage on a subaquous slope. A more likely explanation would be undercutting and sliding/caving in of a channel margin. During the sliding the blocks deformed plastically. After rotation of bedding back to the horizontal, the axes of the largest slump fold trends about 160^{0}/05S. Probably the slump slid along the bottom of the channel floor, thereby the frontal part of it was bent up and folded backwards. Peacock (1965) suggested sliding towards the W. However, from the shape of the slump fold it is here suggested that the mowement was towards the ENE, approximately perpendicular to the fold axes.

Figure 6.38 shows the finer grained, intraformational conglomerate immediately above the channel floor in the westermost part of the outcrop. The channel in the outcrop can be traced for about 60 m, probably it is wider. The floodplain was probably not large enough for free meanders to develop. That is suggested by the restricted amount of overbank material. It is considered more likely that rivers were of low sinuosity, and that the relatively small amount of overbank material is the result of the tendency of low sinuosity streams to comb their floodplains (Allen, 1965).

Palaeocurrent measurements in Member K were extremely difficult to obtain. Whenever trough cross-bedding was present, it proved to be impossible to measure directions on bedding surfaces. At Balsnes there were indications of southeastwards directed currents. This may be consistent with the ENE-directed slide described above. In addition, steep channel margins have been observed to trend in southeasterly and easterly directions, for example the one in Fig. 6.35 and others on the islands west of Balsnes. Usually the overall flow direction is sub-parallel to the steepest margin of a channel, i.e. the meander bend.

When going farther E in the investigated area, overbank sediments occur only sporadically. Braided channels become more important, and the massive bedded sandstones dominate completely. Current directions were not obtained, but N of Bekkvikholmen trough cross-beds indicated current directions to the NW.

Member K was probably deposited in the distal reaches of a system of sandy, alluvial fans. The Balsnes area contains a somewhat larger number of fine grained deposits. It is thought to represent the transition to a floodplain with higher sinuosity rivers. To the east braided rivers dominate (Ch. 7.2).

6.4 BALSNES FORMATION (MEMBER L)

The member is almost entirely made up of Facies 10, conglomerate, but in some ares there are scattered lenses and layers of Facies 5, massive bedded sandstone, usually not more than 0.5 m thick. Occasionally the lenses show Facies 6, horizontal bedded sandstone or Facies 7, trough cross-bedded sandstone (Fig. 6.39).

The Balsnes Formation represents a coarse, alluvial fan deposit. Usually it is too coarse for sedimentary structures to be seen. The high degree of rounding of pebbles, cobbles and boulders shows that the sediments probably are river transported. The high degree of rounding seems to contradict the fact that it is very coarse and massive, and therefore should be deposited on a more proximal part of an alluvial fan. However, if enough time is available, the fragments may become well rounded even if transport is short. There is no evidence of resedimentation, e.g. cobbles composed of lithified sediments, but it is likely that boulders were deposited and redeposited many times during transport.

In addition to transport by rivers, it is considered likely that mass flow transport has played an important role, e.g. debris flows. Such processes may partly be responsible for the homogeneous character of the deposits. River currents must have been very competent to move the big boulders, the largest observed was 2 m across. One possibility is that transport was by debris flows, which later were reworked. Lateral shifting of channels removed the finer pebbles and only the coarser material was left behind. Interstices were subsequently filled by sand, e.g. from stream floods. Sandstone lenses and layers are outwash from the more homogeneous deposits. There are often grain-supported conglomerates at the bottom, laid down as lags in shallow channels. Winnowing has removed the finer material in between. Horizontal bedded sandstone lenses are a result of deposition in the upper flow regime. The sandstone beds exhibit the same coarseness as the sand found in the matrix. Imbrication is poorly developed, and could not be used for current direction measurements. However, Fig. 6.39 shows a weak tendency for imbrication to be present. No other good current direction indicators exist.

In conclusion, the Balsnes Formation is largely an ortho-conglomerate deposited in river channels on the proximal part of a system of alluvial fans. The fans were probably relatively big, since fragments are well rounded even on their proximal parts. The composition of clasts (Table 4.4) supports this assumption, as a large portion of them are extrabasinal. The fans probably shed sediments towards the southeast because:

Along the northern margin the Balsnes Formation lies above sandstones of the Vollan Formation. Along the southeastern margin the Balsnes Formation lies above sandstones and conglomerates of the Vollan Formation. This means that possibly the upper part of the Vollan Formation equates with the Balsnes Formation along the southern margin, being a mid-fan equivalent to the proximal fan of the Balsnes Formation along the northern margin. However, as already mentioned, within the Balsnes Formation there are indications of palaeocurrents flowing towards the northwest. For further discussion and sketches, see Ch. 7.2.

CHAPTER 7

STRATIGRAPHIC CORRELATION TO OTHER "DEVONIAN" DEPOSITS AND PALAEOGEOGRAPHY

STRATIGRAPHIC CORRELATION TO OTHER "DEVONIAN" DEPOSITS AND PALAEOGEOGRAPHY

7.1 STRATIGRAPHIC CORRELATION TO OTHER "DEVONIAN" DEPOSITS

The present author has not visited Fosen and only briefly visited Smøla, and is therefore not able to present detailed correlations. However, after mapping on Hitra some aspects have emerged, which will be compared with previous authors observations.

The first who tried to correlate the O.R.S. on Hitra with other O.R.S. areas was Peacock (1965). From studies of cementation and composition on Hitra and Smøla he concluded that the "Hitra Beds" are probably older than the "Smøla Beds". The "Smøla Beds" are soft and friable and have a clean carbonate cement, while the "Hitra Beds" are highly indurated, and are cemented by epidote, silica and chlorite. He also noted the difference in sedimentary environment. He finally concluded that the "Smøla Beds" are of Lower to Middle Devonian age, while the "Hitra Beds" are older, perhaps even contemporaneous with the Hovin Series (Richter, 1958).

Siedlecka (1975) was the next who tried to correlate the Hitra succession with other deposits. Based on fossils, lithological similarities and degree of deformation she assumed the Vollan Formation to link up with the Austrått Formation on the Fosen Peninsula and the Balsnes Formation to be equivalent with the Kråkøya Granite Conglomerate. To the SW the Vollan Formation was correlated with the Edøy Conglomerate (Fig. 7.1).

It is now clear that the eurypterids found in the Vollan Formation and the Glasøy Conglomerate (Fig. 7.1; Reusch, 1914; Størmer, 1935; Peacock, 1965) existed throughout the Wenlockian, Ludlovian and Lower Devonian, not only during the Ludlovian and Downtonian. Therefore correlations and age determinations previously made between Hitra and Smøla may be incorrect. The lithological similarity between the Glasøy Conglomerate and Member A was mentioned (Siedlecka, 1975). However, the Glasøy Conglomerate is also locally derived, and reflects local basement composition (Peacock, 1965). On this basis they can not be correlated in time. The similarity in composition is only a result of similar basement rocks in the two areas.

A correlation between the Edøy Conglomerate and the Balsnes Formation was made by Siedlecka (1975), as both were claimed to contain large amounts of sedimentary rock fragments derived from earlier deposited "Devonian" rocks. This break was supposed to represent the Downtonian/Devonian boundary. Extremely few sedimentary rock fragments were found in the Balsnes Formation. Neither is there strong evidence for the suggested break in deposition and subsequent erosion at the base of the Balsnes Formation (Ch. 4.9.7). The only clear break in deposition observed is between Members F and G. These points suggest that Peacock's idea of the Hitra Group being older than the O.R.S. southeast of Smøla still may be correct.

If the Balsnes Formation is the time equivalent of some of the sediments southeast of Smøla, one should expect to see a change in either composition, angularity or size where this event occurred in the Smøla succession, because it is supposed to represent a major rejuvenation of source areas (Ch. 7.2). In this way the Balsnes Formation could just as well correlate with the NW Kyrhaug Formation as with the Edøy Conglomerate.

To the NE the Vollan Formation was correlated with the Austrått Formation (Siedlecka, 1975). This was done on the basis of similar depositional environments and similar degree of alteration of the sediments. The assumption does not seem unreasonable. The Balsnes Formation was correlated with the Kråkøya Granite Conglomerate based on the assumtion that both formations contain large amounts of sedimentary rock fragments. The correlation may be correct, but, as already mentioned, there are very limited amounts of sedimentary rock fragments in the Balsnes Formation.

The present author is of the opinion that the most important factor for correlation must be degree of alteration of the sediments. Fossil evidence is not conclusive, neither is composition, as nature of source areas varied from place to place. Therefore the Hitra Group is probably time equivalent with the Austrått Formation, while all other sediments on Fosen are younger. From fossil plants, animals and spores a Lower to Middle Devonian age was obtained for the Storfosna Sandstone (Fig. 7.1 and Allen, 1976).

Beaty (1970) calculated a sedimentation rate of 7.5-15 cm/1000 years in alluvial fan sediments. Similarly, Miall (1978) indicated sedimentation rates varying from 4-50 cm/1000 years for main-late orogenic foreland, interarc, forearc and marginal basins. He concluded that such short term, high sedimentation rates are to be expected, as some areas are uplifted at rates of up to 30 m/ 1000 years. If one assumes the highest depositional rate of 15 cm/1000 years for the alluvial fan sediments SE of Smøla (Beaty, 1970), 25 million years are needed to deposit the 3800 m measured (Beaty, 1970). His highest sedimentation rate is used because the O.R.S. SE of Smøla is very coarse and massive and bears evidence of rapid sedimentation. Large mudflows and debris flows indicate this (author's observations). Miall's (1978) highest rate of sedimentation would give even much more rapid deposition.

If one assumes that the Hitra Group is of Ludlovian age, most of the O.R.S. succession SE of Smøla, which probably is of the same age as the upper formations at Fosen and Storfosna, could still be deposited during the Lower Devonian. It is therefore not unlikely that the Hitra Group and The Austrått Formation are older than the other deposits. As mentioned in Chs. 4.7 and 6.2.2, the presence of bioturbation in the Aune Formation introduces a problem, as continental bioturbation should not be expected in the Silurian. The question of age is therefore still unsolved.

It is not possible to correlate the formations of the Hitra Group with the O.R.S. outcrops on the small islands north of Fosen and Frøya (Ch. 2.4), as very little information is available from those areas. However, from palynological studies Allen (1976) obtained an Emsian age equivalent north of Tristeinen. Therefore these deposits are probably younger than the Hitra Group. No ages have been obtained from the O.R.S. on Froan (Sæbøe, 1972), Tarva and Asenøya (Vogt, 1929) or Inngripen and the skerries to SW of Smøla (Bryhni, 1974). Therefore they can not be correlated in age with the formations of the Hitra Group.

7.2 PALAEOGEOGRAPHY AND DEPOSITIONAL MECHANISMS

The palaeogeography has earlier been a subject of speculation. Peacock (1965) measured the direction of currents from ripples, and concluded that they were flowing from the northwest, suggesting that there was a landmass in that direction. From petrography he stated that the Aune Formation was locally derived, while the Vollan Formation and Balsnes Formation were mainly derived from a source composed of acid and intermediate plutonic, volcanic and hypabyssal rocks. Sedimentary and metamorphic fragments were uncommon. He interpreted all of the Hitra Group to be deposited in an estuarine environment.

Siedlecka and Siedlecki (1972) were the next to speculate on palaeogeography. They did not carry out detailed palaeocurrent measurements, but from petrographic evidence they concluded that the source area had to lie to the SW, in the Smøla area. They suggested a continental depositional environment with lakes, meandering and braided rivers and alluvial fans. The model was further developed by Steel, Siedlecka and Roberts (1985) (Fig. 7.2).

The Hitra Group is here (Ch. 6) interpreted as representing deposits of a variety of sedimentary environments; braided rivers on alluvial fans, braidplains, meandering and low sinuosity rivers and floodbasins with lakes and floodplains. On the islands SE of Smøla deposition was mainly in braided rivers on alluvial fans. Debris flows and mudflows are common (Peacock, 1965; pers. observations). In the Outer Fosen Area Siedlecka (1975) recorded similar processes, that is, deposition by braided rivers, debris flows and mudflows on alluvial fans. The presence of some point bars, channel fills and temporary lake deposits was also suggested. Temporary lakes are documented by the occurrence of dolomites in the Storfosna Sandstone (Fig. 7.1). Dolomites, which were formed in lacustrine environments adjacent to alluvial fans, were also described from Flatskjær, Hustadvika (Bryhni, 1974). The sandy and conglomeratic alluvial fans, which can be traced on the islands towards Smøla, were interpreted as parts of a bajada (Bryhni, 1974).

Observations by the present author show general eastwards flowing currents for the Aune and Vollan Formations. Measurements are based on ripples, cross-bedding, channel directions and syn-sedimentary deformation structures. There are directional variations both towards the NW and SE (Ch. 6). Very few current direction measurements were obtained from the Balsnes Formation.

Conglomerates and sandstones in the Vollan and Balsnes Formations contain large numbers of fragments which are not present in the local basement, e.g. limestone, jasper, chert and acid and intermediate extrusive rocks (Tables 4.2 and 4.4). As already mentioned, these rocks occur on Smøla. To the NE, on northern Hitra and Fosen, the metamorphic equivalents are found (Kollung, 1964). The restricted number of high grade, metamorphic rock fragments in the Hitra Group were probably derived from the north.

The Hitra Group is situated on Ordovician intrusives. This implies that the area was strongly denudated before deposition of the O.R.S. started. Sedimentary rocks of the Hitra Group were either derived from the autochthonous rocks below the Caledonian thrust-nappes (Ch. 3.1), or, if Hitra is part of a thrust-nappe, from the thrust-nappe. The possibility of thrusting during/after deposition of the "Devonian" is discussed in Chs. 2.4 and 11.1.

The present observations of current directions, as with earlier observations, show the weakness of being collected only from the westernmost outcrops of the Hitra Group. It is in no way certain that currents further east were flowing in the same direction. Steel, Siedlecka and Roberts (1985) have for the Vollan Formation suggested a model where meandering rivers in the SW flow to the NE, become braided and finally disappear (Fig. 7.2). Eastwards there is evidence of stronger and more violent currents, now represented by conglomerates and trough cross-bedded and massive bedded conglomeratic sandstone. At the same time there are fewer mudstone sequences. The present author tends to believe that the eastern area was dominated by braided tributaries, which predominantly flowed to the N, S and W towards the central part of the basin (Balsnes area). This central part was dominated by floodplains and lakes, and received finer grained sediments (Fig. 7.3a). From petrographic evidence it is probable that alluvial fans, if situated to the southeast, were either small or situated far away (Fig. 7.3a). They did not contribute significantly to what is now preserved of the Hitra Group.

In the Balsnes area the Vollan Formation is about 960 m thick. Going to the east it thins to about 70 m in the Grindvik area (Ch. 4.9.4). Its members, if they do not wedge out earlier, successively onlap the basement (Plates 1A and 1B; Fig. 7.3b). The succession is probably thinner in Grindvik because this area was a more distal part of the sedimentary basin during deposition of the Vollan Formation. Because the Aune Formation is present all along the area from Aune to Grindvik, it is suggested that the pre-Vollan Formation depositional surface was approximately at the same elevation everywhere. During deposition of the Vollan Formation the central part of the depositional basin (Balsnes/Aune area) was depressed more rapidly than the distal parts, and therefore the central part received a thicker pile of sediments.

The thinner sedimentary succession in Grindvik is an indication that the Hitra Group, at least during deposition of the Vollan Formation, did not link up with the deposits on Fosen (Ch. 7.1). However, it is believed that the original sedimentary basin had a wider extent than what is now preserved, as no typical marginal facies are present along the northern or southern margins. Sedimentary marginal breccias similar to those in the probably fault bounded Devonian Hornelen and Kvamshesten areas (Steel et. al., 1977; Ch. 2.4) are not observed.

The sudden incoming of coarse conglomerates in the Vollan Formation shows that rejuvenation of a highland source area took place during deposition. Uplift of source areas and deposition in basins as a result of movements along marginal faults has been the commonly accepted explanation for all O.R.S. basins in Norway (Steel et al., 1985). Major, marginal faults were also active in other areas of Devonian deposition, e.g. the Midland Valley Graben (Ch. 2.5). In Western Norway possible marginal faults have been described as trending ENE-WSW to E-W (Ch. 2.4; Nilsen, 1968; Steel and Gloppen, 1980). Based on the present outcrop pattern a similar system was indicated in the Outer Trøndelag Region (Siedlecka, 1975; Steel et al., 1985). It is now clear that this pattern is solely a result of the "Devonian" being preserved in structural synclines along the coast of Trøndelag (Ch. 8).

Weather marginal faults were present and active or not during deposition of the Hitra Group is a question open to speculation, but, if present, they were probably situated well away from the present outcrop margins, not in the position indicated in Fig. 7.2. Rejuvenation of highland source areas may very well take place without major fault activity. Examples of very rapid uplift without major faulting were described by Miall (1978) (Ch. 7.1).

The Balsnes Formation was predominantly deposited as alluvial fans, which were built after rejuvenation of source areas. There is no systematic change in maximum or mean maximum boulder size along the length of the area (Table 4.3). If deposition took place towards the ENE or WSW one should expect such a change over a distance of 23 km. Neither is there significant variation with respect to clast composition (Table 4.4).

As mentioned in Ch. 4.11, the composition of the Balsnes Formation is different from that of the lower formations. This may either be due to a continued and deeper erosion of the rocks in the source area of the Vollan Formation, or it may be the result of derivation from a new source area. In Chs. 4.9.7 and 6.4 it was described how the Balsnes Formation apparantly interfingers with the Vollan Formation along the southern margin of the area, and that it therefore probably was derived mainly from the north, northwest and northeast. This conclusion is supported by the presence of high-grade metamorphic rock fragments, similar in composition to rocks in the basement on Northern Hitra and Fosen.

It is possible that there was activity along marginal faults during deposition of the Balsnes Formation. However, such faults were not observed, and therefore nothing more about possible locations or trends can be inferred.

The Balsnes Formation was probably built as an extensive bajada/piedmont formed by the coalescing of alluvial fans along the southern margin of an uplifted, highland source area. Similar interpreatations have been suggested for the conglomerates on Fosen (Siedlecka, 1975) and the conglomerates southeast/southwest of Smøla (Bryhni, 1974).

In summary, all O.R.S. deposits of the Outer Trøndelag Region are probably continental. No clear marine indicators have been observed. This fits well with the model that among others was proposed by Anderton et al. (1979), in which Trøndelag and Western Norway were situated in the middle of the Old Red Continent during Devonian times (Fig. 2.1).

C H A P T E R 8

STRUCTURAL GEOLOGY

8 STRUCTURAL GEOLOGY

8.1 INTRODUCTION

The Old Red Sandstone Hitra Group sediments have been subjected to at least three phases of deformation. The main structure is two large, tight, ENE-striking synclines with undulating axes and subvertical axial planes. Superimposed on the flanks of these are a number of smaller second and third order climbing folds. Along the southern margin there was ductile deformation within shear zones. Later phases of deformation are represented by kink bands, crenulation cleavage, open folds and brittle deformation within shear zones. The synclines are cut by faults trending either sub-parallel to the main axes or at about ninety degrees to these. Joints have about the same trends.

8.2 GENERAL STRUCTURES OF THE MAIN SYNCLINES PRODUCED DURING D1

The general outline of the structure is a series of synclines striking about 65 degrees (Plates 1A, 1B, 2A, and 2B). There does not appear to be one single syncline with a NE-dipping fold axes as reported by Siedlecka and Siedlecki (1972). There is rather a system of "en echelon" folds which die out longitudinally. An example is seen 1.5 km E of Akset (614200 32250), where the main syncline gradually dies out and is replaced by a new syncline 450 m to the SE. In between there is a smaller anticline. This kind of sidestepping is only observed to affect the main synclinal axes at one other place. Southwest of Litlvann (616200 34750) there is a right lateral sidestep of 300 m of the axes before it continues to the ENE. Smaller scale, "en echelon" folds are common along the southern margin.

The strike and dip of axial planes were found to be about 65 degrees/subvertical. This was obtained from plots of strike and dip from various portions of the synclines. From the plots it became clear that the fold axes have an undulating nature, and that the "bigger syncline" can be divided into two large, noncylindrical synclines forming an "en echelon" pattern. The transition between them is the area described E of Akset. The westernmost syncline has been found from Figs. 8.1 and 8.2. Fig. 8.1 shows a plot of poles to bedding planes in the area from Aunåa to Langnes. It gives an almost horizontal axes. Fig. 8.2 shows the same from the Akset area, and gives an axes which plunges 35 degrees to the WSW.

Fig. 8.3 shows a plot of poles to bedding planes in the area from Sandstad (615500 35500) to Litlvannet (616200 34750). The plunge of the axes appears to be gently to the NE, but the figure does not give a clear picture due to later folds being superimposed on D1-folds. In the easternmost part of the area the plunge is again about 30 degrees to the WSW (Fig. 8.4). The centre of the eastern syncline was actually observed in a new road section where road 714 takes off from road 713. The synclinal centre has a V-shape, which was expected from strike and dip measurements on the limbs of it.

8.3 S1-CLEAVAGE AND ITS RELATIONSHIP TO FOLDING

In some areas a strong cleavage has developed during D1, preferentially in the finer grained mudstone beds. It can best be seen in Member F at the coast SE of Furuholmen (610950 21500) (Figs. 8.5 and 8.6). Roberts (1981) was the first to observe and describe the locality. The cleavage is represented by insoluble residues of clay minerals and oxides concentrated on closely spaced, slightly sinuos, sub-parallel fracture surfaces. The cleavage is a transition between a slaty cleavage and a fracture cleavage. As noted by Roberts, there is sometimes a coarse fracture cleavage developed in sandstone beds. In some areas carbonate concretions in mudstones have been rotated into the plane of the cleavage during deformation (Roberts, 1981; Siedlecka, 1977).

There is a slight variation in cleavage directions in this area, but when taking the average of the measurements and plotting it against bedding, a fold axes was found trending $98^0/14$ (Fig. 8.7). This is consistent with what was found using Fig. 8.1. Another plot from Member D on Furuholmen gives a fold axes trending $246^{0}/04$ (Fig. 8.8). From Member J a plot of measured bedding cleavage intersection lineation (Fig. 8.9), kink fold axes and constructed bedding cleavage intersection lineation gives an average axes of about $230^{0}/30$ (Fig. 8.10). Both of these are consistent with the general trend. The variation is a result of the noncylindrical nature of the main syncline (Ch. 8.4.1).

In addition to the cleavage described , there is in some areas an even stronger cleavage developed. This can be observed on Selnes (610800 25200), Mamnavollen (612250 28200), Hamn (612500 28900) and Akset (613000 29500), in other words the places where the "Devonian" is most strongly deformed (See below). There cleavage is developed in mudstone/sandstone interlayered sequences in second order synclines, and can by eye be seen as fractures spaced 5-10 cm apart trending sub-parallel to the axial planes of second order folds (Fig. 8.11). These fractures develop along the long limbs of third order climbing folds. Sometimes cleavage is so predominant that folds can hardly be recognized.

Cleavage is best developed in the western and southwestern area, and will be further described and used later.

8.4 MINOR FOLDS PRODUCED DURING D1

Contemporaneously with development of the main synclines during D1, smaller scale, climbing folds were superimposed on the limbs of them. They only occur in the westernmost half of the area, and are mainly present on the southern limb.

8.4.1 Selnes, a type area

For closer examination of these folds Selnes was chosen due to its good exposures (Fig. 8.12). The style of folding is representative for D1-folds in other areas as well. This also applies to the directions of axial planes and fold axes.

Both second and third order climbing folds are present on the southern limb of the main syncline. On Selnes the second order folds do not reach an amplitude of more than 20 m. Other places they are larger. Maximum recorded amplitude for third order folds is 1 m, but normally does not excede 15 cm. As for the main synclines, the higher order folds are noncylindrical. Second order folds on Selnes have a maximum plunge of 10 degrees, in other areas it may reach 25 degrees. Third order fold-plunges vary from 0 to 90 degrees. Individual third order fold axes may curve in plunge through arcs of as much as 130 degrees (See below). Azimuth directions are fairly constant, arcs of curvature were not observed to be larger than 20 degrees.

A plot of fold axes from Selnes gives a clear trend. They all fall within a girdle outlined by two small circles. Average strike of the axes become 063^0 (243^0) with a deviation of 18 degrees to both sides (Fig. 8.13). An axial planar, spaced slaty cleavage is present (Fig. 8.11), and is especially well developed in synclines of second order folds (Fig. 8.12). The average cleavage direction calculated is $063^0/90^0$, which also is the axial planar orientation (Fig. 8.13). This is the same as the axial plane directions for the main synclines.

Most fold axes, belonging to D1, plot within the girdle outlined by the two small circles on Selnes. There is a spread of plunges of different fold axes from plunging weakly to the WSW through the vertical to plunging weakly to the ENE, that is, a spread of plunges through an arc of nearly 180 degrees. Nowhere else was such a strong variation in plunge directions of different fold axes observed at one locality. Within other limited subareas, e.g. at Hamnavollen, the majority of third order folds are cylindrical in nature, and exhibit a relatively high degree of constancy in axial orientation. The spread at Selnes results from the high degree of noncylindrism of third order folds in the second order synclines, where deformation has been most intense. On horizontal surfaces folds could be seen as ellipses with the fold axes plunging in opposite directions at the ends (Fig. 8.14). The fold type varies from normal and inclined with varying plunges to vertical and reclined folds. Most of the folds have convergent axial plane cleavage fans, and belong to Class 1C (Ramsay, 1967), but there are also similar folds (Class 2). Similar folds were observed in thin sections, while 1C-folds could be seen on a 10 cm scale in the field. The folds usually form kink bands (Fig. 8.15).

On the north-dipping limb of the southernmost second order syncline on Selnes it was possible to see how deformation becomes more intense as one approaches the centre of the syncline. The synclines between noncylindrical anticlines are frequently less regular and not as well formed, their form being dictated by the space requirements of the dominating anticlines. It was nowhere possible to observe the axes of folds for more than 30 cm along the strike, but the wavelength of axial curvature usually appeared to be more than four times the profile wavelength. In the plunge culminations the curvatures of the axes are often smooth, while in plunge depressions it may be much tighter and sharper. This is the same tendency as observed by Ramsay and Sturt (1972) from Sørøy.

There is an increase in the number of third order parasitic folds as one approaches the centre of second order synclines, and the limbs of folds become more stretched as they tighten up. From the S towards the centre of the southernmost second order syncline (Fig.8.12) one goes from virtually undeformed sediments to sediments where it is very difficult to discern bedding. This is partly due to fewer sand layers in the centre of the syncline, a phenomena which leads to less competency during deformation. However, it proves that it is bedding and not an earlier cleavage that is folded, as might be suspected from microscopic studies. As mentioned by Roberts (1981), there is a crude preferred orientation of phyllosilicates parallel to the primary lamination because of earlier sediment compaction (Ch. 9.1). The strong deformation in the centre of the syncline makes it difficult to discern this fabric from what could have been an earlier cleavage. There is also a strong development of bedding-parallel stylulites pre-dating D1. They came into existence during compaction, and are present in mudstone beds throughout the succession. This will be further discussed in Chapter 9.1.

8.4.2 Hamnavollen

At this locality some of the most well formed D1-folds are preserved. The general picture are two second order anticlines climbing to the south (Fig. 8.16). The amplitude is about 60 m for the northernmost and somewhat larger for the southern one. On the northern limb of the syncline in between a number of well formed folds can be observed. They are often arranged in kink bands (Fig. 8.17) with amplitudes of folds between 5 cm and 10 cm. In addition there are some larger folds with amplitudes of about 0.5 m (Fig. 8.18). A plot of fold axes, poles to kink bands and poles to axial planes (Fig. 8.19) shows that these structures have the same trends as the structures on Selnes. Fold axes fall within the girdle defined on Selnes (Fig. 8.13).

In the central part of the syncline a slaty cleavage has again formed. It has the same orientation as on Selnes (Fig. 8.13). The cross-section on Fig. 8.16 shows the general outline of the fold pattern at Hamnavollen. In general the folds plunge about 20 degrees to the ENE.

8.4.3 Havnøy (610000 23000)

The southwestern portion of this island is a syncline with fold axes plunging about 70 degrees to the ESE. The axes flattens eastwards. North of this there is an anticline with weakly northwestwards plunging fold axes striking in the same direction. The northern limb is overturned. The fold pattern is rather complex, probably because of the proximity to the shear zone bordering the "Devonian" to the south (Ch. 8.9.3). Amplitudes of the folds are in the order of 75-100 m.

In the centre of the syncline a conjugate fold has developed with fold axes sub-parallel to the larger fold (Fig. 8.20). Chevron and box folds were also observed in Members J and K on the coast NW of Balsnes. A plot of third order fold axes and axial planes from the centre of the anticline shows a wide spread of the data (Fig. 8.21). Some of the measured folds were polyclinal folds. There is a predominance of reclined and plunging inclined folds at this locality.

8.4.4 NE of Kalvhaugterna (610300 24000)

This area comprises the most complex deformation pattern in the area of investigation. The succession is strongly folded and later faulted. This, together with poor exposures made it difficult to map out the structures, but the most important synclines and anticlines are indicated on Plate 2A. A peculiar structure has formed just to the north of Kalvhaugterna. An anticline to the north is followed by a syncline to the south, and then comes a bigger anticline with an amplitude of at least 150 m and a steeply southwards dipping axial plane. The anticline had originally a shape as a cone. A southeastwards dipping, listric, normal fault has later down-faulted the southeastern half of the cone to give the peculiar outcrop pattern. A cross section is shown on Plate 1B.

8.4.5 Promontory NE of Bekkvikholmen

The only place where Member B crops out strongly folded is at the promontory NE of Bekkvikholmen and on the small island to the west of it (610750 24750). Both second and third order folds are present. The single second order fold present is only weakly developed, the southern limb of a small anticline is approximately horizontal. However, on this limb a number of tight, parasitic M- and Z-folds are existent. They have amplitudes up to 15 cm, and exhibit a wide range of directions for fold axes and axial planes as seen on Fig. 8.22. This is partly a result of later disturbance, e.g. due to the emplacement of an intrusion (Ch. 10), but the noncylindrical shape of the folds plays an important role. Fold axes have a clear trend NE-SW as expected. The axial planes are steeply inclined to the NW. Under the microscope cleavage could be seen to be axial planar to D1-microfolds (Fig. 8.23).

8.4.6 Hamn, Akset and E of Akset

All the way along the coast from Hamnavollen to the SE of Terningvann there is an extensive development of parasitic second order folds. On the western part of the peninsula at Hamn there are two anticlines with a syncline in between. The northern is the bigger with an amplitude of about 75 m. Going to the E, the southern anticline disappears, while the northern one grows bigger. The folds at Hamn have plunges of about 30 degrees to the WSW. Probably some of them link up with the ENE-plunging folds at Hamnavollen in a noncylindrical nature. A plot of third order parsitic folds from Hamn is shown in Fig. 8.24. Axial planes dip steeply to the NW. A cross-section from Hamn is given on Plate 18.

At Akset NE of Hamn one single overturned anticline dominates the picture. The axial surface has a dip to the NNW of about 75 degrees, and is slightly nonplanar. In deciding the form of the strucure a combination of way up criteria and vergence of smaller parasitic folds had to be used. Few such folds could be observed, but their axes were constructed from bedding cleavage intersection lineation and gave a clear picture (Fig. 8.25). The average plunge is about 30 degrees to the WNW. The amplitude of the larger anticline is in the order of 100 m, but westwards it rapidly dies out.

Across the bay E of Akset there is another set of folds. A larger, northern anticline may possibly be the continuation of the anticline at Akset (Plate 18).

8.4.7 Balsneslangvann

Only at one place along the northern limbs of the main synclines were parasitic folds observed. The locality is situated on the southern shore of Balsneslangvann, about 100 m to the W of the small ness (611750 24400). A plot of the measurements shows that fold axes plunge 15-20 degrees to the ENE (Fig. 8.26). Axial planes dip about 60 degrees to the NNW. Amplitudes of folds were not bigger than 10 cm. Going eastwards from this locality, fold amplitudes steadily increase to about 1 m. At the same time plunges increase to about 60 degrees with unchanged azimuths of the fold axes. Further eastwards the folds become smaller again, the plunge of fold axes come closer to the horizontal, and finally, about 500 m to the E of the first mentioned locality, no more parasitic folds were observed.

The pattern described here is consistent with the noncylindrical nature of folds described from the southern limb of the westernmost, main syncline.

8.4.8 West of Sandstad

About 1 km west of Sandstad there is a large, open anticline followed by a smaller syncline to the south. In this smaller syncline the Balsnes Formation has been preserved, while elsewhere it is eroded away. The strike of the syncline is parallel to the D1-trend. The strike of the anticlinal axes changes from ESE-WNW to E-W as one go to the W. The fold axes changes from being ESE-plunging to being horizontal. For the syncline there is an opposite trend; there is a change from horizontal to an east-northeasterly plunge as one goes to the west.

8.5 DISCUSSION

The style of folds belonging to the D1-generation is clearly noncylindrical in nature, though within smaller sub-areas they may be cylindrical. On a larger scale the noncylindrism is demonstrated by the plots of strike and dip for the main synclines. The larger parasitic folds are shown to be noncylindrical from plots at Hamnavollen and Hamn. There is a plunge depression of the fold axes between these two localities. Minor parasitic folds are noncylindrical as explained from Selnes. Within very small sub-areas folds may appear to be cylindrical, e.g. at Balsneslangvann, but when traced laterally, folding can also there be seen to be noncylindrical in nature.

Wavelength and amplitude of the folds are clearly dependent upon lithology, being smaller for thin bedded and finer grained sediments. This is also the case with axial curvature. On a small scale two types of noncylindrical folds were observed. The two are plane noncylindrical and nonplane noncylindrical folds. Both may either remain constant along strike or tighten/widen. The latter category is asymmetrical, and there may even be transitions to disharmonic folds, as can be seen at the coast in Member J. This suggests modification by later stage flattening, in which the maximum extension occurred parallel to the layering of the long limbs, causing rotation of the mid limb to unfolding in one direction and to a reclined position in the other (Ramsay, 1962). However, disharmonic folds are only weakly developed.

Fold axes always curve within the trace of the axial surface. This, together with the similarity in style and axial orientation, suggests that the noncylindrism is primary and not a consequence of refolding. Within different sub-areas strikes of axial planes are fairly constant. However, dip may vary by up to 30 degrees away from the vertical both to the north and to the south. That is probably a result of nonplane noncylindrism for some of the second order parasitic folds as well. The genesis of noncylindrical folds was discussed by Ramsay an Sturt (1972).

On Selnes there is a transition from the normal congruous style of folding into aberrant folds. Third order parasitic folds on the relatively gentle anticlines have fold axes approximately parallel to the second order fold axes. The axial surfaces make an angle of about 90 degrees to the mean layering. As one approaches the steep limb towards the synclines, this angle becomes acute. The axial curvature of the minor parsitic folds become more intense, and the variation in plunge direction of different fold axes increase. There is also an increasing variation in azimuths of fold axes and degree of non-cylindrism. The significance of the angle between mean layering and axial surface for the variation in plunge direction of fold axes was described by Ramsay and Sturt (1972). There is no transition into real incongruous folds. A folding relationship similar to the situation on Selnes could be observed at the promontory NE of Bekkvikholmen.

Everywhere, all through the area, there are variations in dip of bedding of up to 30 degrees over relatively short distances. This can only be explained by parasitic folds and variations in the form of the main synclines, as there is no evidence of ramping or thrusting. Rotated fault blocks are not responsible for the variatons. Few such blocks were detected, and they do only occur along the southern margin. All in all the structures described indicate that D1 has been a relatively strong deformational event.

8.6 D2-STRUCTURES

Roberts (1981) described a system of NNW-SSE trending kink bands with consistent westerly downstep in the area from Selnes and westwards. This phenomena was also observed by the present author (Fig. 8.27). Few kink bands were observed, however, figure 8.28 shows a plot of their poles along with some measured fold axes. As can be seen, the kink bands strike at an angle of about 90 degrees to D1-folds. On the southern limb of the main syncline just to the N of Steinbitholmen (610250 23600) fold axes plunge approximately 40 degrees to the NW. On the northern limb, at Furuholmen, the plunge is 25-40 degrees to the SE. Amplitudes of the kink folds are in the order of 5-10 cm. At the coast NW of Balsnes, in Member J, there is a set of monoclines with amplitudes up to 40 cm and the same westerly downstep as for the kink bands described. Axial planes have a dip of 50-80 degrees to the E. Fold axes plunge about 40 degrees to the S (Fig. 8.29 and Fig. 8.30).

The fact that the plunge changes from S to N as one goes across the centre of the main syncline may indicate that these structures have been refolded. However, if the kink bands and monoclines are later than the main deformation, the influence of already tilted bedding may have been so strong that fold axes had to occur in the plane of bedding.

500 m to the W of Litlvannet there is another set of folds affecting the earlier structure. A system of gentle, open, weakly southeastwards plunging synclines and anticlines have folded the succession in the centre of the main syncline. Strikes of these folds are approximately perpendicular to the D1-fold. A similar system was observed SW of Bekkviktjern (611400 25250) with gently northwesterly plunging, open

folds. On Selnes some good examples of refolded isoclinal and tight folds were seen. Fig. 8.31 shows one of these with an axial plane trending $148^0/31E$, which is consistent with directions observed for the kink bands (Fig. 8.28).

The author is of the opinion that both the kink bands, the monoclines, the folds SW of Litlvannet and Bekkviktjern and the refolded folds on Selnes belong to D2. The strongest evidence for this are the refolded folds at Selnes (Fig. 8.31). The variation in fold axes orientation from N-S to NW-SE is probably a result of variation in the strike and dip of bedding on the main D1-fold. This phase of deformation was probably not a very strong one. No cleavage was observed which can be associated with it.

8.7 CRENULATION CLEAVAGE

The strongest deformation phase affecting the Hitra Group was D1. This was subsequently followed by another, weaker deformation, D2, generally without cleavage development. However, a crenulation cleavage has been formed in finer grained sediments along the southwestern coast. The age relationship between the crenulation cleavage and the described D2-structures is still a question open to speculation. Nowhere the two phenomena were observed at the same locality. Therefore it is possible that the crenulation cleavage is older than the D2-structures, but it can not belong to the same deformational event, as a different stress regime must have acted (Ch. 11.1).

The crenulation cleavage is hardly recognizable in the field, only at Hamnavollen and Selnes it was observed directly. Figure 8.19 shows a plot of the crenulation cleavage measured at Hamnavollen. The dip is approxiamately 75 degrees to the SE. At Selnes the dip was about 55 degrees to the N (Fig. 8.32). Generally this cleavage strikes in an E-W direction. Roberts (1981) detected this later, north-dipping crenulation cleavage at Selnes. However, from the two plots available it appears to vary around the vertical. When examined in thin section the crenulation cleavage can readily be observed as micro-kink bands affecting the earlier, axial planar, slaty cleavage (Figs. 8.33 and 8.34; Chs. 9.2.2.4

and 9.2.3).

Sometimes not only a crenulation cleavage, but a new slaty cleavage has formed. In a thin section from Selnes it was possible to see how an earlier, tight D1-fold with axial planar, slaty cleavage has been refolded with development of a new, axial planar, slaty cleavage. The angle between the two is 20-30 degrees (Fig. 8.35).

A sample from the centre of the second order syncline at Hamnavollen was strongly brecciated. Fragments of mudstone exhibited isoclinal folds with a well developed S1 axial planar, spaced crenulation cleavage. What is crenulated are bedding-parallel stylolites (Ch. 9.1). Then there has been brecciation and recrystallization. About 50% of the rock is now made up of calcite, but the mudstone fragments are little rotated with respect to each other. A later cleavage with an angle of about 20 degrees to S1 is especially well developed in the calcite matrix. It can be seen as a strong, preferred, long axes orientation of calcite grains (Fig. 8.36). For further discussion of the crenulation cleavage, see Ch. 9.2.3.

8.8 OTHER LATE STRUCTURES

In the eastern part of the area from Badstuvik to Grindvik there is a set of folds which probably originated contemporaneously with the crenulation cleavage. The direction of dip of the bedding varies from NW to W, and even to the WSW. The amount of dip is constant at about 30 degrees. Fold axes constructed from strike and dip of bedding plot in the manner shown in Fig. 8.37. Lack of bedding measurements made it difficult to trace the fold axes westwards, but there is a tendency for them to bend into parallelism with the main syncline, and graduallly they die out. In the eastern part of the area the axes plunge to the NW, farther to the SW they plunge to the SW before going back to the NW. The structures are shown on Plates 1B and 2B. This deformational event is probably partly responsible for the wide spread of third order D1-fold axes at Havnøy and at the promontory NE of Bekkvikholmen.

8.9 FAULT PATTERN

8.9.1 Introduction

Throughout the area many more or less important faults were detected. The dominating fault directions are ENE-WSW and NW-SE (Fig. 8.38). It is usually easy to see which faults displaces others, but to tell how old the oldest are, is worse. Some of the faults may be reactivated pre-Caledonian or Precambrian lines of weakness. Roberts (1983) mentioned a set of NE-SW trending syn-sedimentary faults. The present author has not observed any such faults. The statement appears to be based on the assumption that there were active marginal faults to the N and S of the depositional basin (Ch. 7.2). However, S of Storvågen (617900 36850) the basal conglomerate suddenly stops to the E against what possibly is a pre-sedimentary fault escarpment. To the W of this line the thickness is in the order of 150 m. To the E of it there is no development of the basal conglomerate. Syn-sedimentary faulting was discussed in the chapter on palaeogeography (Ch. 7.2).

8.9.2 Reverse Faults

To the SW of Litlvannet (615550 33500-616000 34250) there is an area 900 m long and 175 m at the broadest where the basement rocks pierce through the sedimentary pile. The lithology is mainly granodiorite and granite, which is similar to local basement composition to the N of the unconformity. On top of this basement wedge some meters of Member A are preserved followed by the Vollan Formation. Below it lies Member B on top of Member A (Plate 1B).

Only at one small outcrop the contact between Member B and the overlying basement wedge could be observed. It showed probably to be a high-angle, bedding parallel reverse fault orientated $060^0/65S$. It was impossible to get a sample from the fault contact itself, but a sample of Member B mudstone taken 5 cm below the contact showed a high degree of cataclastic crushing and pressure solution. It was set through by pseudotachylitic veins in many different directions (Fig. 8.39). The laminae were strongly affected by micro-folds. Half a metre away from the fault the mudstone did not appear to be deformed at all.

The diorite above the fault is a cataclasite. A thin section of a sample taken immediately above the fault showed very strong recrystallization with large amounts of secondary epidote and chlorite. All mineral grains are crushed (Fig. 8.40), some are intruded by veins of pseudotachylite, which penetrate the rock in many directions. The diorite is heavily crushed for at least 1 m above the contact.

A possible model for the faulting is shown on Plate 1B, cross section E. The diorite is supposed to represent the sliced off top of a palaeohill. For this model a movement of at least 300 m along the fault is required. Whether or not there is a strike slip component involved is difficult to say. It was not possible to see the lateral continuation of the fault. Therefore it was terminated against the NNE-SSW trending fault to the W and in Member B to the E. Probably it continues beyond these points. It is suggested that movement took place during D1.

It is also possible that this feature originated as a thrust prior to folding, as bedding-parallel movement can more easily take place when bedding is horizontal/subhorizontal. The lack of ductile deformation shows that if this is a thrust, it was probably of minor importance. A possible model is shown in Fig. 8.41.

8.9.3 Trondheimsleden Shear Zone along the Southern Margin

The largest and undoubtedly most important fault/shear system in the area is the main shear zone trending sub-parallel to the coast along the southern margin of Hitra. On Jøssenøy south of Sandstad strongly sheared and partly mylonitized basement rocks similar to those described W of Kallarnes can be observed(Ch. 3.3.3; Fig. 3.10). These rocks are here interpreted as the result of a major, post depositional movement along the Trondheimsleden Shear Zone. It is most likely post-depositional because:

1. No rock fragments of metadiorite/ mylonite have been found in the Hitra

Group.

2. "Devonian" sediments to the N of the sheared basement rocks W of Kallarnes are similarly strongly sheared and deformed (Ch. 3.3.3).

Along the southern margin of the area on Steinbitholmen (610250 23600), Kalvhaugterna (610350 24000), the promontory NE of Bekkvikholmen and on Langnes (611250 26400) there are outcrops of virtually undeformed basement rocks similar in composition to basement rocks observed along the northern margin. On the promontory NE of Bekkvikholmen there is an unconformity to the overlying sediments. The other places there are fault/shear contacts as described in Ch. 3.3.3. Some of these faults are probably secondary splays related to the main shear zone, which probably continues farther to the S. They all trend ENE-WSW, and are grouped along the southern margin of the area.

The major shear zone trends about 060^0. The deformed diorite can be traced from Jøssenøy, across Aunøya before it reappears on Vedøya to the E. On the small islands between Vedøya and Grindvik the diorite is undeformed, implying that the shear zone lies to the S of them. The westernmost exposure of deformed diorite is on an island 2 km to the W of Jøssenøy (612350 32200).

Even if the 2 m thick cataclasite on Kalvhaugterna represents an important shear zone, degree of deformation is not comparable with the strong, ductile deformation experienced within the main shear zone to the S. The cataclasites on Kalvhaugterna and Steinbitholmen have a dark green colour because of a high content of secondary chlorite and epidote. The rock is set through by a number of calcite-filled veins, which were folded during progressive deformation. The matrix of the cataclasites is to fine grained for compositional studies, however, calcite, quartz, epidote and chlorite could be recognized.

Some of the faults, e.g. the one on Kalvhaugterna, now appear to be normal faults with downthrow to the N. This can be due to later deformational events. However, it is not uncommon to have reverse movements along a vertical fault. The orientation may also have changed during progressive deformation.

Deformation along the Trondheimsleden Shear Zone probably took place during more than one phase. The strongest event is represented by ductile deformation and mylonitization. Either thrusting, transpressional shear movement or a combination of these were responsible for the deformation and metamorphism within the shear zone. Another, weaker deformation is expressed by brittle faults and cataclasites (See also Ch. 8.11). They probably originated during transpressional movements.

It is not likely that all the faults in the area suffered a compression during deformation. Some may have been higher order splays with components of normal faulting.

On seismic sections from Trondheimsfjorden to the south of Hitra (Bjerklia, 1983) a set of listric, normal faults trending ENE-WSW could be recognized. The faults exhibit downthrow to the NW in the order of at least 200 m, and in half-grabens sediments of a possible Mesozoic age are preserved (Bjerklia, 1983). This probably indicates that at some time during the Mesozoic a tensional tectonic regime operated. Normal faults originated either along older lines of weakness or as new features. The mechanisms which operated during deformation are discussed in Ch. 11.1.

<u>8.9.4 Transverse Faults</u>

Along the length of the area there is a group of faults trending transverse to the main synclinal axes. They have highly varying orientations, but two populations seem to dominate, one trending about 105^0 and another 155^0 (Fig. 8.38). Most of the faults are sub-vertical. The faults are either normal or dextral strike-slip faults, but some examples of sinistral strike-slip faults exist, e.g. the one offsetting basement onto Hitra W of Kallarnes (Plates 1A, 1B, 2A and 2B).

Generally, there is less deformation along these faults than along those paralleling the main shear zone. However, along the fault trending NW-SE at Langnes there is a several metre broad, heavily brecciated zone, but no cataclasite has been produced. A similar, 10 m broad zone of fault breccia was observed along the E-W trending boundary fault at Grindvik (Fig. 8.42).

Fig. 8.43 shows an example of a small horst produced by normal faulting. Normal faulting can also be observed E of Balsnes, where the Balsnes Formation has been downthrown against the Vollan Formation. The fault plane dip 33 degrees to the east. A 2 cm thick zone of flinty crush has developed along it.

Few of the transverse faults were observed to be offset by NE-SW trending faults, the opposite is usually the situation. Therefore most of them are belived to belong to the latest fault generation.

8.10 JOINT PATTERN

All the way along the southern margin there is an intricate pattern of jointing. Mesoscopic joints are especially abundant in areas with well developed parasitic folds, or where the main Trondheimsled Shear Zone passes close by. Orientation of sub-vertical, mesoscopic joints was measured at five different localities along the coast. The results are given in Figs. 8.44a-e. Figs. 8.44a and 8.44b show two important orientations, one ENE and one WNW to NNW. Fig. 8.44c shows one predominant direction E-W, and another minor one N-S. Figs. 8.44d and 8.44e have a dominant direction NW-SE.

The observed pattern of mesoscopic jointing is consistent with one group paralleling the strike of folds at the locality, and another group being approximately transverse. This is the same general tendency as for the faults. In the conglomeratic centres of the main synclines, e.g. at Aksethusberg (613750 29800), the sediments are strongly fractured by strike-parallel, sub-vertical joints (Fig. 8.45). This indicates that at least the strike-parallel joints bear a relationship with folding (Hobbs et al., 1976).

In addition to the mesoscopic joints there are two sets of macroscopic joints, one trending ENE-WSW and the other NNW-SSE. These joints are of the same type as those described by Nilsen (1968) from the Solund Devonian. They are generally sub-vertical, show a maximum separation of adjacent walls of up to 10 meters, and individual joints can be traced for several kilometers. The ENE-WSW-striking joints show the largest lateral separation. Because these joints are so clearly defined, they are probavly of a late origin. They are possibly related to the Late Cenozoic uplift of the Scandinavian landmass (Holtedahl, 1960; Nilsen, 1968).

8.11 PSEUDOTACHYLITES AND CATACLASITES

Parallel to bedding there are occasionally some green lenses and bands of a fine grained rock. This could be studied in the road cut in the Balsnes Formation N of Bekkviktjern (611450 25000), where they have developed in a coarse grained sandstone. The lenses have a maximum thickness of 1-2 cm, and a length of 5-20 cm. In addition there are thread-like, very thin, undulating bands and strings of the green material.

When examined under the microscope the greenish brown, fine grained rock proved to be pseudotachylite, which has later devitrified. Floating within the fine grained matrix are grains of strongly deformed quartz and minor amounts of epidote, chlorite, feldspar, sphene and rock fragments. Feldspars are brittly deformed, quartz grains are partly recrystallized (Fig. 8.46 and Ch. 9).

The pseudotachylites probably originated as a result of bedding-parallel, flexural slip during folding (Hobbs, Means and Williams, 1976). Locally this process produced temperatures high enough for melting of the rock. Pseudotachylites were also noted in connection with the reverse fault/?thrust SE of Litlvann (Fig. 8.39), in sediments close to the main shear zone NE of Kallarnes and in the diorite and metadiorite along the southern margin. In an analogous way high temperatures resulted during shear movement.

At the coast at Balsnes, about 30 m above the base of Member K (610650 22200), there is something that at the first glance appears to be a half a metre thick bed of horizontal bedded sandstone. Laterally it is continuous for at least 20 m. Closer examination showed that it is not very dissimilar from the pseudotachylites at Bekkviktjern. However, none of the very thin, thread- like, undulating bands and strings could be observed. Under the microscope the rock proved to be heavily crushed (Fig. 8.47). Some zones are close to an ultracataclasite (Higgins, 1971).

Floating within the groundmass are brittly deformed feldspars, zircons, sphenes, quartz grains and epidote crystals. The groundmass is mainly recrystallized microcrystalline quartz plus some epidote. Penetrating in different directions are veins filled with bigger quartz crystals and epidote. In the most finely crushed portions of this sample cleavage was visible , a feature not commonly seen in sandstones, not even with a microscope. This proves that the cataclasite is older than the last formed cleavage. Probably the cleavage is S1, as the crenulation cleavage has not been observed this far away from the main shear zone anywhere else.

The bedding-parallel cataclasite probably originated as a result of flexural slip during folding. However, the temperatures did not get high enough to melt the rock, as has been the case N of Bekkviktjern. Cataclasite is a common phenomena in association with faults throughout the area. Some were described in the section on faulting, so this will not be repeated again.

C H A P T E R 9

DIAGENESIS AND METAMORPHISM

9 DIAGENESIS AND METAMORPHISM

9.1 EARLY COMPACTION

Roberts (1981) was the first to report on a crude, preferred orientation of phyllosilicates paralleling the primary lamination. Based on Siedlecka's (1975) assumption that 2-2.5 km of "Devonian" sediments had once overlain the coastal districts of Norway, he concluded that it represent a compactional fabric. The presence of a strong compactional fabric was also noted by the present author. It is especially observed in mudstones belonging to the Vollan Formation. They have suffered a strong compaction, with resultant concentration of residual, insoluble minerals, probably mainly clay minerals and oxides, as stylolites on bedding parallel surfaces (Figs. 9.1 and 4.6). The stylolites are very thin, usually much less than the tenth of a millimetre. The length may vary from the tenth of a millimetre to several centimetres. They can only be examined under the microscope. However, beds with large amounts of mud contain a number of them. The stylolites are rarely perfectly straight. They rather have an anastomosing nature, where single stylolites may join and split laterally, thereby wrapping around "lenses" of coarser siltstone (Fig. 9.1). These siltstone "lenses" appear to be enriched in quartz cement. The most fine grained laminae are completely dominated by minute, anastomosing stylolites. Most of the quartz and feldspar appear to have been dissolved during compaction.

As mentioned above, there is a preferred, bedding-parallel orientation of phyllosilicates, especially white micas (Fig. 9.2). It is assumed that these were reorientated during the compaction which produced the stylolites. Other minerals lying in contact with the phyllosilicates, e.g. quartz, show a high degree of pressure solution on bedding-parallel surfaces. Sutured contacts between minerals are common. It is difficult to estimate how volumetrically important the primary compactional pressure solution has been because of later pressure solution during deformation.

However, the extensive development of stylolites in the mudstones suggests that primary compressional pressure solution led to a significant thinning of at least some of the mudstone sequences. Within mudstone beds sand diapirs were observed (Fig. 6.25), of which some are folded. If it is assumed that they originally were approximately straight and perpendicular to bedding, some are shortened by up to 50 %. This may partly have taken place during deformation, but may also be the result of early compaction.

A factor inhibiting pressure solution was early cementation. Within calcite concretions Siedlecka (1977) showed the existence of nearly uncompacted, sand-filled burrows (Fig. 4.5). She suggested that there are two generations of calcite concretions. The early generation shows a stronger compactional fabric outside than within the concretions. The later generation shows approximately the same degree of compaction both within and outside the concretions. During deformation the concretions were rotated. Their long axes, from being sub-parallel to bedding, have been rotated towards the S1-cleavage plane. Some of the smallest concretions were rotated into complete parallism with S1 (Roberts, 1981). For further discussion on this subject reference is made to Siedlecka (1977) and Roberts (1981).

9.2 METAMORPHISM

9.2.1 Introduction

Little work has earlier been carried out on metamorphism in the Norwegian O.R.S., mainly because the sediments were previously generally regarded as unmetamorphosed, however, Sturt (1978) mentioned that some areas have been affected. Metamorphism was not a major objective of this thesis, but the following is an attempt to figure out how strongly this process has affected the Hitra Group. It is difficult to divide the metamorphism into separate phases, therefore it will be discussed in one section. It appears to be clear that D1 was the strongest deformational event, and metamorphism during this phase affected the whole area. During later deformational events smaller sub-areas along the southern margin were subjected to metamorphism.

9.2.2 Mineralogical changes in the sediments

The following is a description of the main mineralogical changes which have taken place in the sediments during deformation and metamorphism. The finer grained lithologies proved to be most valuable for this purpose.

9.2.2.1 Quartz

Quartz has recrystallized during deformation of the sediments. This is best seen in the finer grained mudstones and siltstones, where grains are elongated parallel to the axial planes of F1-folds (Fig. 9.3). These quartz grains always show undulose extinction as a result of deformation. Certainly the quartz grains also showed undulose extinction before deformation of the "Devonian", as they were mainly eroded from rock complexes which have undergone Caledonian deformation (Ch. 3.1). The grains were strained, resulting in crystal lattice dislocations, and thereby undulose extinction (Spry, 1969).

Elongation of quartz crystals is the result of pressure solution-recrystallization and plastic deformation. It is difficult to estimate the degree of plastic deformation, as most crystals showed undulose extinction prior to erosion/deposition. However, elongation did probably take place during deformation. Pressure solution and recrystallization were certainly important processes. Pressure solution has been strongest at the most mobile boundaries. Weakly developed pressure shadows behind quartz and feldspar grains were observed in areas where the deformation has been very strong, e.g. NW of Kallarnes (Ch. 3). In these "shadows" very fine grained, granular quartz has grown. The crystals are relatively unstrained and randomly orientated. This show that they grew during a late stage of deformation.

It is not uncommon to find granoblastic quartz in the sediments. It usually occurs as small, polycrystalline aggregates of rather equidimensional crystals. Within some of the aggregates, the crystals only show a weak undulose extinction. These probably crystallized during a late stage of deformation. It is very common to find quartz crystals which have grown over other minerals, especially over minute micas and epidotes (See below).

The quartz grains have a stronger alignment parallel to S1 when there are large quantities of clastic mica present. This is a result of extensive pressure solution at quartz-mica interfaces (Fig. 9.4). The micas occur between the quartz grains instead of growing over them. Therefore the shape fabric of quartz is a result of pressure solution along grain boundaries (Spry, 1969). Especially along the limbs of folds in tight/isoclinally microfolded mudstones, sub-parallel S1, there is a strong preferred orientation of quartz grains (Fig. 9.5). The large number of micas parallel to the foliation inhibited grain boundary migration of quartz. Therefore the dimensional preferred orientation of quartz grains mainly resulted from growth parallel to the foliation (Spry, 1969). A number of small, unstrained, new crystals with this orientation supports elongation as a result of growth.

In the hinge region of folds the tendency for a preferred orientation of quartz crystals is much weaker, and new, relatively strain-free crystals have grown. At the grain boundaries between clastic micas and quartz grains new, minute, white micas have grown sub-parallel to the boundaries. They can only be observed under a microscope with high magnification. When there is little clastic mica present, the tendency for a preferred orientation of quartz grains is much weaker, as the tendency for gliding along quartz grain boundaries is reduced (Spry, 1969).

Cross-cutting the bedding in different directions are quartz-filled veins. They contain unstrained crystals, which probably grew after the main deformational events.

9.2.2.2 Feldspar

There is a varying degree of alteration of feldspars, from nearly unaltered to completely retrograded grains. This partly represents a varying degree of alteration prior to erosion and deposition, but it is clear that feldspars suffered strong retrogradation during deformation and metamorphism. The most common products of feldspar retrogradation are fine grained white mica, possibly sericite, epidote, zoisite and calcite. These minerals, except for zoisite, were observed to have crystallized in the matrix, and in cases of very strong retrogradation it is impossible to see what is matrix and what originally were feldspar grains. Sericitized feldspars contain large numbers of minute sericite needles scattered throughout and randomly orientated within the crystals. Saussuritized plagioclases have got a dirty appearance because of new epidotes.

Pressure solution of feldspars has taken place. Many feldspars are elongated parallel to F1-axial planes, especially when they lie in contact with muscovites (Fig. 9.4). In the pressure shadows of quartz and feldspar grains minute, new feldspars were observed. In the pressure shadows growth has also taken place on the parent crystals. The parent crystals are retrograded, while new growth rarely shows retrogradation. The grain boundaries between old crystals and new growth are outlined by dirty rims. Fractured feldspar grains are often "healed" by new feldspar growth. One example of a recrystallized K-feldspar was observed in a quartz filled micro-fracture. The grains in the fracture are unstrained, and have therefore crystallized after the main deformational event. Plastically bent feldspars with undulose extinction are common in areas with strong deformation. If temperature was low during deformation, the strain-rate must also have been low to get plastic deformation of feldspars (Higgins, 1971). However, temperatures were probably relatively high (Ch. 9.2.2.4).

9.2.2.3 Epidote

Epidote is among the commonest new minerals that have grown in the Hitra Group succession, and is found both in sandstones and mudstones. It is usually a very fine grained alteration product of feldspars, now found within what must be termed matrix. However, along the southern margin it was observed as clusters of equidimensional or prismatic, up to 3 mm long, well formed crystals without any preferred orientation. It also occurs as undeformed crystals filling small joints and veins. These undeformed crystal were formed after the main deformation of the Hitra Group.

9.2.2.4 Chlorite and White Mica

Along with quartz and epidote, chlorite is the volumetrically most important new mineral in the Hitra Group. It is the common reaction product during retrogradation of biotites and feldspars. Most of the amphiboles in the sediments have also been replaced by chlorite. Biotites are in all stages of alteration, from nearly unaltered to completely replaced by chlorites.

Within the finer grained lithologies of the Hitra Group there is a large number of minute, white mica needles, which appear to have grown during metamorphism. They are especially abundant in the microfolded mudstones along the southern margin of the area. They are also common in the pressure shadows of larger detrital quartz and feldspar grains, where they have grown with different orientations. Most commonly the (001)-planes are sub-parallel to S1.

The intention was to analyze the white micas on a microprobe to get their exact composition. This could again tell which temperatures and pressures which operated during their formation (Velde, 1967). However, because of the small size of the needles, it was not possible to confine the electron beam to the minerals being investigated only. The results were poor, and could not be used for compositional determination. Similar analyses have been carried out on new, white micas from the O.R.S. in Kvamshesten (Western Norway) and in the O.R.S. south of Smøla (Sturt et al., 1986). There the new micas proved to be phengites, probably formed at temperatures of about 400^{0}C (B.Sturt, pers. comm.). When using these data on Velde's (1967) diagrams, pressures of 3-4 Kb were obtained. It is possible that the minute, white micas in the Hitra Group are phengites too, but this remains to be documented. They may very well be some kind of mixed-layer mineral. If that is correct, it would indicate temperatures lower than 400^{0}C during metamorphism (Pique, 1982).

Within the isoclinally microfolded mudstones on Selnes there appear to be chlorite-white mica aggregates, but the amount of chlorite is restricted. There is a very strong alignment of both white mica and chlorite (001)-planes parallel to S1. The aggregates have been deformed by the later crenulation cleavage (Fig. 9.5). Kink bands parallel to the crenulation cleavage are common in the white micas, and the micas have partly been rotated into parallelism with the crenulation cleavage. The micas are highly deformed, which either suggests that they are of detrital origin or that they were formed early during deformation. Chlorites, on the other hand, exhibit a varying degree of deformation, suggesting continuous solution and growth. The aggregates have formed following splitting along the (001)-planes of white micas, where chlorite has grown. This process was described by v.d.Pluijm and Kaars-Sijpesteijn (1984) and Woodland (1985) as an early metamorphic process.

Some chlorite-muscovite-biotite aggregates of detrital origin were found in the sandstones. They are very similar to aggregates in the dioritic basement in Hestvika. It is therefore not impossible that the aggregates in the sediments originated prior to erosion and deposition. The aggregates in the mudstones are to small and vaguely defined to say anything about orientation variation between white mica and chlorite from fold hinges to fold limbs. That could tell exactly when chlorites have grown (v.d.Pluijm and Kaars-Sijpesteijn, 1984).

9.2.3 Discussion

Pressure solution was widespread during deformation and metamorphism of the Hitra Group. On a large scale this can be envisaged from sutured contacts between pebbles and cobbles in the conglomerates (Fig. 9.6). Under the microscope sutured contacts and elongation of grains parallel to S1 can be demonstrated to be widespread (Figs. 9.3 and 9.4)

Where pressure solution was most intense, and there is an appropriate lithology, a cleavage has developed. As mentioned earlier (Ch. 8.3), the cleavage is outlined by dark seams of insoluble clay minerals and oxides concentrated on fracture surfaces. Accompanying this is an elongation of minerals between the fracture surfaces, either as a result of rotation, plastic deformation or pressure solution/recrystallization processes.

S1 varies from a coarse fracture cleavage in sandstones to a stronger, slaty cleavage in mudstones. The cleavage in mudstones varies from non-planar spaced through planar spaced to a planar penetrative cleavage (Pique, 1982), of which the last mentioned represents the strongest deformation.

In tight/isoclinally microfolded mudstones without clear fracture surfaces, there has been extensive concentration of white mica along the limbs of folds. All feldspars have been retrograded, and much of the quartz has gone into solution. This is in line with the series in order of decreasing mobility for low grade metamorphic rocks established by Gray and Durney (1979), in which calcite, quartz and feldspar should dissolve first. White mica is relatively immobile. In veins and fold hinges quartz and calcite has recrystallized. As silica was lost, this process probably in places led to a significant thinning of the original sedimentary succession (Beach, 1979).

Also along with development of the crenulation cleavage there was widespread pressure solution. The cleavage is outlined as dark seams of residual, insoluble minerals, and is only found within mudstones. The seams are continuous only for short distances, e.g. 1-2 mm. At the ends there are transitions into microfolding before it disappears. Such crenulation-fold bands may be up to 10 cm long. The folds may have wavelengths up to 1 mm. Gray and Durney (1979) and Gray (1979) stated that both microfolding and solution-deposition processes, in that order, are necessary for crenulation cleavage development. In other words, the crenulation-folds predates the crenulation cleavage. There does not appear to have been new growth of flaky minerals parallel to the crenulation cleavage. The cleavage is only a result of transfer of dissolved species from the limbs to the hinges of microfolds.

It is very difficult to estimate how strong metamorphism was during burial and deformation of the sediments of the Hitra Group. As mentioned by Winkler (1979), there is no clear transition between diagenesis and metamorphism. Many minerals are formed during diagenesis which also occur during metamorphism, e.g. feldspar, chlorite and quartz. The strong deformation and cleavage development along the southern margin indicates that metamorphism has taken place. Very low-grade minerals such as zeoliths were not observed, possibly because they disappeared during advanced stages of metamorphism. Neither were zoisite nor clinozoisite, which occur at the transition from very low-grade to low-grade metamorphism (Winkler, 1979), observed. It is probable that the sediments were subjected to low-grade metamorphism. However, no diagnostic mineral which could give any definitive information about pressures or temperatures was observed.

CHAPTER 10

INTRUSIVE ROCKS

10 INTRUSIVE ROCKS

10.1 INTRUSIVE ROCKS

Only at one locality in the "Devonian" on Hitra was an intrusion recorded. This was at the small promontory 150 m NE of Bekkvikholmen. Siedlecka and Siedlecki (1972) concluded that it was of pre-"Devonian" age. This was based on the assumption that also the intruded sediments were of pre-"Devonian" age. However, it is now clear that these sediments belong to Member 8 (Ch. 3.3.3). Therefore the intrusion is of "Devonian" age or younger.

The intrusive body is about 20x3 m in size. It consists of quartz, K-feldspar, plagioclase, some perthite and accessory sphene. Along grain boundaries there are minor amounts of secondary chlorite. Generally the feldspar is strongly retrograded. The intrusion has a pink colour, is medium to coarse grained, and can be classified as a granite or a leucogranite.

On the southern side there is a 5 cm thick, greyish, chilled margin preserved. This contact cuts straight through the strongly folded sediments, and makes an angle of 10-20 degrees to the strike of folds (Fig. 10.1). The orientation of folds does not change as one approaches the contact. The intrusion is probably younger than D1, because:

1. The intrusive contact cuts straight through the deformed sediments without being deformed at all.

2. If the intrusion was older than D1, one should expect such a rigid body to influence directions and style of folding close to the contact. This was not observed. The contact cuts straight through folds.

The exact age of the intrusion is still unknown. In a thin section from the intrusion no zircons were observed. However, a sample was analyzed on a X-ray spectrometer. It yielded 62 ppm Zr, which is enough for U-Pb radiometric dating if enough sample material is available. Radiometric dating was not carried out during this project, however, it would be very useful both for dating of the sediments and their deformation.

Farther south, in the Kristiansund Area, Pidgeon and Råheim (1972) dated some granitic dikes and pegmatites. Rb-Sr whole-rock and mineral isochron ages from 385 to 393 +/- 7 m.y. were obtained. After correcting their Rb^{87} decay constant from 1.39×10^{-11} y^{-1} to 1.42×10^{-11} y^{-1}, the ages become about 377 and 385 +/- 7 m.y. (Middle Devonian). These results were interpreted as the igneous expression of the Caledonian metamorphism in the Basal Gneisses. Because the boundaries between the geological epoches have been revised, these ages may now represent an early Svalbardian event. It is therefore not unlikely that the intrusion in the Hitra Group also took place during the Svalbardian Orogeny (early Upper Devonian), but after the main deformational episode (See above).

According to Thirlwall (1981) Caledonian subduction probably continued until Middle Devonian times when the final continental collision occured (Ch. 2.4). If this is correct, the intrusion may be related to a late, final closure of Iapetus. Thirlwall (1981) suggested a similar origin for granites south of the Southern Uplands Fault.

In the Devonian of Solund there has been extrusion of rhyolitic and trachytic lavas (Ch. 2.4), possibly indicative of a continental rift environment. The age of the volcanics is not exactly known, but is probably early Middle Devonian (Furnes and Lippard, 1983). It is not impossible that these volcanics belong to the same generation of igneous activity as the intrusion in the Hitra Group. Middle and Upper Devonian intrusive activity was also widespread in Scotland, on Shetland and in SE Greenland (Ch. 2).

A Permian age was obtained from a syenite porphyry dike at Tustna NE of Kristiansund (Råheim, 1974). Similarly, a possible Permian age of an ultrabasic biotite lamprophyre has been recorded from Ytterøy in

Trondheimsfjorden (Råheim, 1974). The compositions and textures of these are rather different from the granite dike in the Hitra Group. Therefore the dike on Hitra is not thought to belong to the same generation, but it may still very well be of a younger age than Upper Devonian.

CHAPTER 11

CONCLUSIONS ON DEFORMATION HISTORY

11 CONCLUSIONS ON DEFORMATION HISTORY

11.1 SEQUENCE OF DEFORMATION AND MECHANISMS

After deposition but prior to deformation a strong bedding-parallel compaction took place in the Hitra Group sediments. This is evidenced by stylolites, pressure solution, preferred orientation of micas and differential compaction in connection with calcite concretions (Ch. 9.1). A considerable burial involving incipient metamorphism of the sediments must have occurred to give this compaction.

The first deformation episode, D1, folded the sedimentary pile into two tight, large, noncylindrical synclines arranged "en echelon" to each other and which trend ENE-WSW. This synclinorium can be traced for at least 160 km along the length of Trondheimsfjorden (Chs. 2.4 and 11.2). On the flanks tight to isoclinal, noncylindrical, parasitic folds developed. The main S1-cleavage was formed during this phase of deformation. Movement along the reverse fault/?thrust SW of Litlvannet and the main shear within the Trondheimsled Shear Zone also probably took place during D1. Other faults were certainly produced, but it is difficult to say for sure which belong to D1.

A compressional tectonic regime with σ_1 directed NNW-SSE is suggested during this phase. That would account for the direction of folds. Possible faults would be those trending E-W and NNW-SSE. Vogt (1928) was the first to ascribe this deformation event to the "Svalbardian Orogeny" in Early Upper Devonian (Frasnian) times (Ch. 2.4). This timing still seems to be correct. See Chapter 2.4 for discussion of the term "Svalbardian".

When movement along the Trondheimsled Shear Zone started is still not exactly known. Probably there were movements both prior to deposition, contemporaneously with D1 and later. Ziegler (1978) suggested that

NE-ENE-trending features such as the Trondheimsled Shear Zone originated as dextral splays from a sinistral mega-shear in the North-Atlantic Region (Storetvedt, 1973; Kent and Opdyke, 1978; Van der Voo and Scotese, 1981) during the Lower to Middle Devonian. This has recently been shown probably to be wrong (Ch. 2.4).

Based on a NW-SE to NNW-SSE extensional tectonic regime and the (probably incorrect) assumption that a sinistral mega-shear and dextral splays existed, Steel and Gloppen (1980) suggested a transtensional tectonic regime during deposition of the O.R.S. in the Hornelen and Kvamshesten Devonian areas and extension during deposition of the Solund Devonian. This model fits well with the suggestion of rift volcanism in the Solund Devonian (Furnes and Lippard, 1983; Ch. 2.4). It fits rather poorly with subduction (compression) suggested by Thirlwall (1981) (Chs. 2.5 and 2.6), even though local transpression was inferred along the northern margin of Hornelen (Steel and Gloppen, 1980; Ch. 2.4). An argument against the model involving subduction until Middle Devonian times (Ch. 2.3) is the lack of marine Devonian sediments in Scotland. Marine deposits are usually associated with subduction zones. Even though the bulk may have been consumed in a subduction zone, some remains should be preserved.

Which models are correct still remains to be seen, however, no signs of syn-sedimentary tectonic deformation, for example associated with movements along the Trondheimsled Shear Zone, were observed in the Hitra Group. Only the incoming of coarse conglomerates indicates tectonic movements in the region during deposition (Ch. 7.2). Therefore the model suggested by Steel and Gloppen (1980) cannot be adopted to explain the tectonic regime during deposition of the Hitra Group. See also Ch. 7.2.

The Svalbardian Orogeny took place during Frasnian times (Vogt, 1929). The present author is of the opinion that the main deformation and metamorphism along the Trondheimsleden Shear Zone occurred at this time, as later deformations were much weaker than D1. It is possible that a large scale, compressive, strike-slip movement (transprssion) occurred, but the very strong, ductile deformation is an indication of thrusting. In Upper Devonian thrusting has been documented from other areas. In Eastern Greenland (Ch. 2.2) basement slices were thrusted through the Devonian succession, and along the southern margin of the Solund Devonian normal faults were converted into thrusts (Ch. 2.4; Nilsen, 1968). From the Devonian in Western Norway recent investigation has shown thrusts cutting the unconformity below the sediments. The thrusts were folded during progressive deformation, or, more likely, during later deformational phases of the "Svalbardian Orogeny" (Ch.2.4) (Sturt, 1983; Sturt, 1984).

As in Western Norway, the Trondheimsled Shear Zone cuts into the unconformity below the sediments (Ch. 3.3.3). The orientation of the mylonitic foliation in the basement varies, but is usually sub-parallel to the orientation of bedding in nearby sedimentary rocks. If bedding is rotated back to the horizontal, the foliation close to the sediments dip weakly to the northwest. This suggests, as in Western Norway, that shearing occurred during folding, and that progressive deformation or later deformations folded the foliation. Such faults as the one southwest of Litlvannet (Ch. 8.9.2), if formed early during deformation (Fig. 8.4), were probably also folded. From what is mentioned above, it appears to be likely that a major component of thrusting was present during movement along the Trondheimsled Shear Zone, and that the Hitra Group is parautochthonous or allochthonous together with the basement to the north.

The later D2-kink bands, monoclines, refolded folds on Selnes and open folds with axes trending N-S to NW-SE (Ch. 8.6) are results of a compression E-W to NE-SW. The downstep of kink bands to the W made Roberts (1981) to believe in a relationship with either Permo-Jurassic rifting or Tertiary uplift of Scandinavia. The latter possibility is considered unrealistic, as the refolded folds on Selnes indicate a stronger, compressive deformation. It is possible that these structures are related to Permo-Jurassic rifting, but they may as well be the result of a late "Svalbardian" deformation.

The late folds in the area from Badstuvik to Grindvik (Ch. 8.8) fit well with a model involving N-S transpression. The Trondheimsled Shear Zone would then be sub-parallel to a plane of no finite longitudinal strain, and folds with axes approximately E-W would develop. The other plane of no finite longitudinal strain would be represented by the NW-SE to WNW-ESE-striking dextral faults cross-cutting the area. This implies that late movements along strike-slip faults within the Trondheimsled Shear Zone had to be sinistral. A strong indication of N-S compression is the E-W-striking crenulation cleavage (Ch.8.7).

This deformational episode was strong enough to cause refolding and crenulation of an earlier cleavage, and is probably responsible for the brittle deformation along faults within the Trondheimsled Shear Zone (Ch. 8.9.3). Accompanying movement along brittle faults, pseudotachylites were formed.The N-S compression is compatible with the stress regime during Permo-Triassic rifting (Ziegler, 1978).

The latest phase of deformation is represented by faults trending NNW-SSE and NNE-SSW, both normal and strike slip faults. Strike slip faults are predominantly sinistral. The faults have offset the main Trondheimsled Shear Zone with up to 200 meters (Ch. 8.9.4).

Late, macroscopic joints have an orientation similar to late faults, and are probably related to Late Cenozoic uplift of Scandinavia. They originated during a strongly tensional tectonic regime.

When the intrusion on the promontory N of Bekkvikholmen occurred is a question open to speculation, but it is clear that it postdates D1 (Ch. 10.1).

No D3 nor D4 have been suggested in the sequence of deformation, as the age relationship between the described D2-structures and the crenulation cleavage/E-W-trending folds is uncertain. Possibly are the D2-structures younger than the last mentioned, but on Selnes there was an indication that the crenulation cleavage overprints a D2-fold. Therefore the eastablished sequence of deformation is probably correct.

It is difficult to deduce the metamorphic state of the sediments, as diagnostic minerals are not present. The strong deformation and cleavage development along the southern margin indicate that some metamorphic mineral reactions must have taken place, of which growth of minute, new, white micas is believed to be one (Ch. 9.2.2.4). The Hitra Group suffered low-grade metamorphism during the suggested thrusting in lowermost Upper

Devonian times.

11.2 DISCUSSION AND CONCLUSIONS

The structures in the Hitra Group are not very complex, neither are structures in other O.R.S. deposits in the Outer Trøndelag Region. Little information is available from the other areas, but the large scale structures are known. On the islands SE of Smøla the succession is folded in a large syncline trending ENE-WSW. Only the northern, overturned limb can be observed (Peacock, 1965). Similar observations have been made on the islands at Hustadvika (Bryhni, 1974). The dip varies from steeply dipping to the ESE to being vertical.

From the Outer Fosen Area there is somewhat more information available (Siedlecka, 1975). The succession is deformed in two NE-plunging, open synclines, which, prior to faulting, were one and the same structure. There is a system of NW-SE and NE-SW trending steep faults and joints. On the islands of Froan to the NW of Fosen Sæbøe (1972) recorded a possible northeast-southwesterly strike of bedding with a steep dip to the SE.

The structures described compare closely with the large D1-structures in the Hitra Group, and probably belong to the same deformational event. The ENE-WSW fold trend is predominant throughout, probably also on the islands north of Frøya and Fosen. The pattern of faulting and jointing is also similar.

From Røragen Roberts (1974) described this "Svalbardian" (Røragian) sequence of deformation: 1. N-S compression with development of slaty cleavage and large folds. Then there was extension with a foliation produced within pelite beds. 2. NW-directed compression with development of mesoscopic folds plunging to the SW and axial planes overturned to the NW. There was contemporaneous thrusting. 3. Normal faulting. 4. N-S shortening with production of kink bands trending ESE-WNW.

The structures in the Røragen Area are less easily comparable with those in the Hitra Group. Point 2 seems to compare with the D1-structures, while

Point 1 seems to compare more closely with the crenulation cleavage trend on Hitra. However, these areas are located well away from each other, and the sedimentary basins probably had a different origin and orientation (Roberts, 1974) that may have influenced their deformation. In addition there were local variations, e.g. close to larger faults/shear zones. As an example, the crenulation cleavage in the Hitra Group is only developed close to the main Trondheimsled Shear Zone, where deformation is strongest.

The change in structural trends in the O.R.S. is evidenced from Western Norway , where fold axes dominantly strike E-W. Going to the NE there is a change to a ENE-WSW direction.

The structural trends in the Hitra Group compare well with structures observed in other "Devonian" areas in the Outer Trøndelag Region. It is more difficult to relate structural trends in the Hitra Group with structural trends in other O.R.S. regions. It is anyway clear that deformation took place under similar tectonic conditions.

From the Surnadal Syncline Råheim (1977) dated some biotites from schists in the lowermost part of the Røros Group (Gula Group). After correction of his Rb^{87} decay constant (Ch. 10.1), they gave ages of about 376-390 m.y. (Middle Devonian). He concluded that these ages either represent later deformations or the final cooling during post metamorphic uplift after the Caledonian Orogeny (See below).

From Stadlandet, which is situated not far away from Devonian outcrops, Lux (1985) carried out isotopic dating in the rocks of the Precambrian basement. His results showed that the area cooled through the blocking temperature of hornblende (500^0C) about 410 m.y. ago, following high grade, Caledonian metamorphism. From biotites he concluded that post-metamorphic cooling through 300^0C occurred about 370 m.y. ago, and that the terraine suffered a linear cooling rate of 5^0C/m.y. This cannot be correct, as the temperature at the surface during deposition of the O.R.S. must have been approximately 0^0C. As examplified by Miall (1978), burial and uplift of terraines can occur very rapidly. It is therefore suggested here that between 410 and 370 m.y. ago there was rapid uplift, erosion, and deposition of the O.R.S. occurred. The biotite age probably shows that the area was deeply buried again just prior to 370 m.y. ago (Frasnian times). Low grade metamorphism and "Svalbardian" deformation probably took place in the O.R.S. successions at this time (Ch. 2.4).

From what is said above, it appears likely that Råheim's (1977) ages of 376-390 m.y. from the Gula Group (See above) represent later deformation, possibly of "Svalbardian" age.

This may indicate that the Surnadal Syncline was affected, perhaps even formed, during an early stage of the "Svalbardian Orogeny" (early Upper Devonian). The structural trends throughout Trøndelag are close to the structural trends in the "Devonian" in the Outer Trøndelag Region. The "Svalbardian Orogeny" may even be responsible for the thrusting in the Ringerike Sandstone of the Oslo Region (Ch. 2.4), of which the stratigraphically upper parts probably are time equivalent with the lower parts of the Hitra Gruoup. This supports Sturt's (1978) suggestion that the lowermost Upper Devonian "Svalbardian Orogeny" (Solundian/Røragian Orogeny) may have had a stronger influence on the structures throughout Southern Norway than what has been assumed until now.

C H A P T E R 12

CONCLUSIONS

12 CONCLUSIONS

12.1 CONCLUSIONS

The Old Red Sandstone Hitra Group on Hitra was deposited ater the Scandian phase of the Caledonian Orogeny in Middle-Upper Silurian times, following a period of rapid post-metamorphic uplift and crustal stretching. Sedimentation took place in a large depositional basin during the Upper Silurian to Lower Devonian. Rejuvenation of highland source areas, possibly along marginal faults, resulted in episodic progradation of coarse, conglomeratic units into central parts of the basin, which were dominated by finer grained clastics. Unconformities are not present within the Hitra Group; probably there was fairly continuos deposition throughout the time interval represented.

The sedimentary succession is of continental origin throughout, no clear marine indicators were observed. Concerning depositional environments for the different members, the following conclusions were reached by studying facies and facies associations:

Member A: Basal breccia, talus and alluvial cones.
Member B: Floodplains and ephemeral streams on distal alluvial fans.
Member C: Braided rivers on middle portions of alluvial fans.
Member D: Lake on a braidplain.
Member E: Braided rivers on middle portions of alluvial fans.
Member F: Low sinuosity rivers on distal portions of alluvial fans and on floodplains.
Member G: Mid/proximal alluvial fans.
Member H: Floodplain/temporary lake in distal reaches of alluvial fans.
Member I: Mid/proximal alluvial fans.
Member J: Lake on floodplain.
Member K: Braided rivers on mid/distal alluvial fans in the east, low

sinuosity rivers/braided rivers on distal alluvial fans/ floodplains in the west.

Member L: Proximal alluvial fans.

After deposition of Members B, F and K there were major rejuvenations of source areas, and coarse wedges built into the sedimentary basin. Deposits of the Aune and Vollan Formations were predominantly shed towards the NE, E and SE, while the Balsnes Formation was deposited towards the SE, S and SW.

After deposition the sedimentary succession was deformed during several phases of deformation, of which the first, the "Svalbardian Orogeny" in lowermost Upper Devonian times, was the strongest. The Hitra Group was folded into two, large, open synclines trending ENE-WSW with superimposed, smaller folds on their limbs. The general style of folding is noncylindrical in nature. In the finer grained lithologies there was locally isoclinal folding. A strong cleavage developed under low-grade metamorphic conditions. Within the Trondheimsled Shear Zone there was ductile deformation, most likely as a result of thrusting, and the whole area suffered strong faulting and jointing. The low-grade metamorphism and ductile deformation within shear zones indicate that the area was deeply buried during D1. It is belived that Hitra, and the Hitra Group, are partly allochthonous, and were originally, prior to thrusting, situated farther to the northwest. A granitic intrusion was emplaced syn/post D1.

The next deformation, D2, is only weakly expressed as kink bands, monoclines and open folds trending NW-SE.

Then came another, rather strong deformation, with crenulation cleavage development and open folding in an east-west direction. Along the southern margin there was brittle deformation within sinistral shear zones, and pseudotachylitic veins were intruded in different directions. The structural style of this deformation indicates that it is of Permo-Jurassic age.

Finally, the area was transected by a set of normal and sinistral strike-slip faults trending NNW-SSE. The age of these is not known.

Macroscopic joints trending ENE-WSW and NNW-SSE are probably related to Late Cenozoic uplift of Scandinavia.