Early post-deglaciation shorelines and sea-level changes along Hardangerfjorden and adjacent fjord areas, W Norway

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Dr. scient. thesis

PREFACE AND ACKNOWLEDGEMENTS

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Paper 1:

Helle, S.K., Anundsen, K., Aasheim, S. & Haflidason, H. 1997: Indications of a Younger Dryas marine transgression in inner Hardanger, West Norway. *Norsk Geologisk Tidsskrift 77,* 101-117.

Paper 2:

Helle, S.K. 2004: Sequence stratigraphy in a marine moraine at the head of Hardangerfjorden, western Norway: evidence for a high-frequency relative sea-level cycle. *Sedimentary Geology 164,* 251-281.

Paper 3:

Helle, S.K., Rye, N., Stabell, B. & Haflidason, H. Manuscript: Early postglacial relative sea-level changes in the mid Hardangerfjorden area, western Norway. (manuscript to be submitted to *Boreas*).

Paper 4:

Helle, S.K., Rye, N., Stabell, B., Prøsch-Danielsen, L. & Hoel, C. Manuscript: Neotectonic faulting and the Late Weichselian shoreline gradients in SW Norway. (manuscript submitted to *Journal of Geodynamics*).

INTRODUCTION

1. Background

Several studies have in the recent years focused on the deglaciation history of Hardangerfjorden and the surrounding region (e.g., Helle et al., 1997; Mangerud, 2000; Bondevik & Mangerud, 2002, Helle, 2004; Lohne et al., 2004; Bakke et al., 2005; Romundset, 2005) (Fig. 1). The rather provocative and controversial article of Helle et al. (1997), where the established glacial chronology for Hardangerfjorden for the first time was questioned, probably contributed to draw interest to the fjord and its surroundings. The general consensus of opinion at the time was that the fjord was ice-filled during the Younger Dryas (YD) (e.g., Aarseth & Mangerud, 1974; Holtedahl, 1975; Andersen et al., 1995). In opposition to this, Helle et al. (1997) proposed that the fjord became deglaciated during the Bølling/Allerød interval and that major parts of the fjord remained ice free during the YD. An important argument in this respect was the evidence for a relative sea-level (RSL) rise, correlated with the YD transgression (Anundsen, 1985).

Although the deglaciation (and sea level) history of the fjord is currently still under debate (e.g., Mangerud, 2000), the high research activity in the region in the recent years has undoubtedly given us new, important information about the Late Weichselian and early Holocene period, regardless of what stand one chooses to take in this debate. The results are important in constraining the dimensions and chronology of the Scandinavian ice sheet, and in constraining glacial rebound model parameters. Ultimately, this has implications for our understanding of the underlying mechanisms that control crustal movements, sea-level changes and landscape evolution.

From a historical perspective it is interesting to see that several of the ideas put forward in the present thesis were called upon also by the early generations of geoscientists. Kaldhol (1941) and Undås (1944), purely on a morphostratigraphic basis, correlated several of the moraines belonging to the Eidfjord-Osa Moraines in inner Hardanger (Fig. 2) to the Ra Moraines in south-eastern Norway, implying ice-free conditions in major parts of Hardangerfjorden, as in Oslofjorden, during the YD. The idea of a marine transgression in the innermost fjordhead areas of western Norway is not of recent date either. Kaldhol (1941, p. 54) described an outcrop section in a marine terrace located close to the marine limit in Lærdal, at the head of Sognefjorden (Fig. 1). The upper unit in this section, 4-m-thick, erosive-based, and consisting of gravel-rich horizontal layers, was interpreted to be deposited during what he called 'landsenkning' (land lowering), i.e., during a RSL rise. In other words, Kaldhol (1941) postulated that a marine transgression had occurred at the head of Sognefjorden and that the development of the local marine limit was closely associated with this event. Neotectonic faulting is another subject that early attracted the attention of the geoscientists working in the Hardangerfjorden area. Reusch (1888) described topographic scarps on the peninsula between Hardangerfjorden and the head of Bjørnafjorden (Fig. 1) which he suspected to be the effect of postglacial faulting. Cone et al. (1963) described a major drop in the fjord sediment level in the mid Hardangerfjorden area which they suspected to be related to postglacial faulting.

Prior to the works of Helle (1993) and Helle et al. (1997), hardly any sea-level records from lake basins in Hardangerfjorden existed (except for the isolation basin data from Bømlo, at the inlet of the fjord; Fægri, 1944; Kaland, 1984; Fig. 1). In previously published shoreline diagrams (e.g., Hamborg, 1983) (Fig. 3), the individual marine terraces and delta terraces along the mid and inner parts of the fjord were correlated on a morphostratigraphic basis. Neotectonism was not taken into consideration in the interpretation of the shore-level data. The correlations were based on the assumption that the highest marine shore features (i.e., the marine limits) were developed during the initial entry of the sea at the time of deglaciation (and thus represent the oldest shore levels), and that the age of the marine limits in general decreases towards the head of the fjord, reflecting a step-wise retreat of the ice front. Hence, the altitudinal distribution of the marine limits was directly
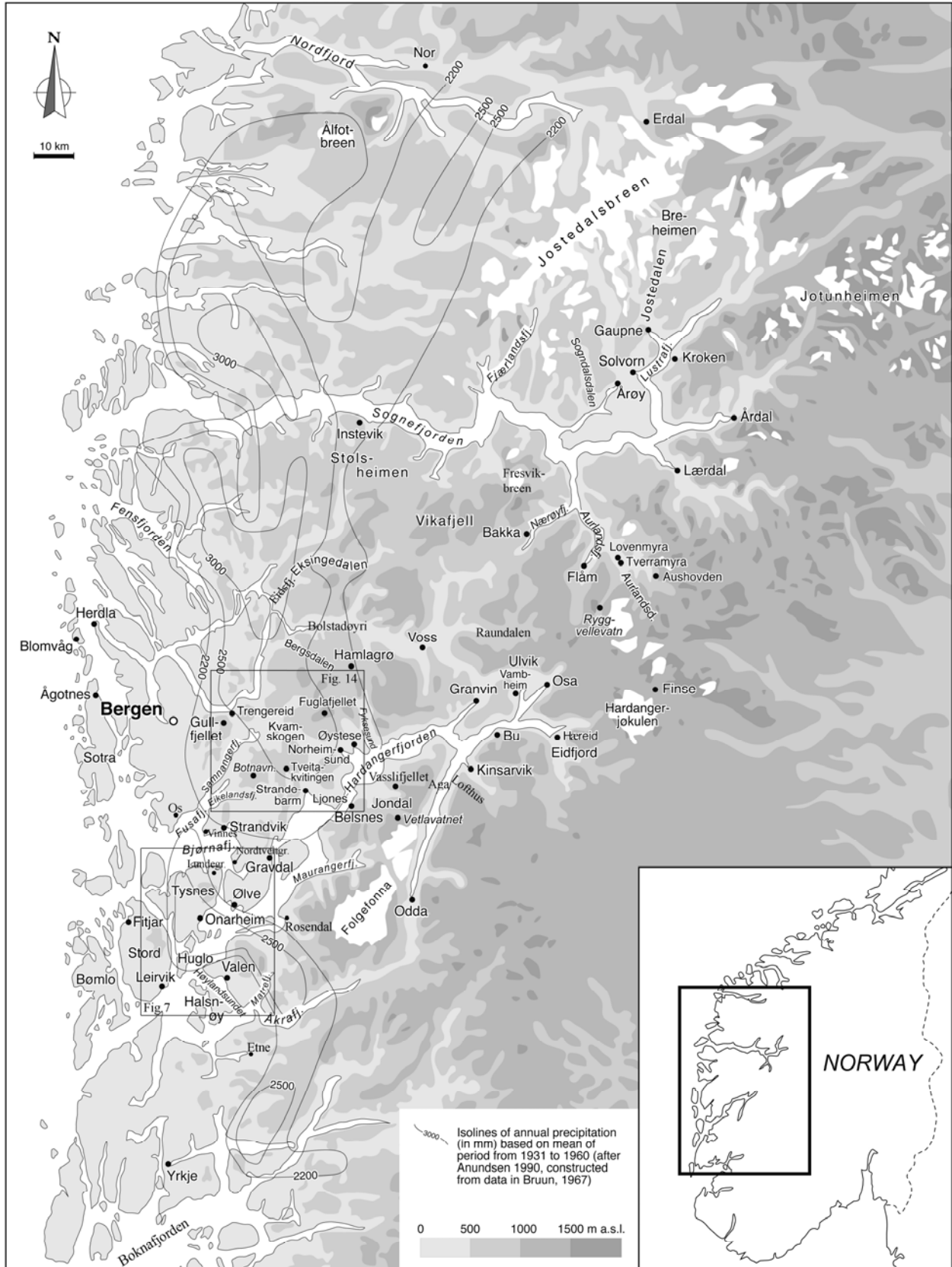
Fig. 1. Location map. Isolines of the present-day annual precipitation are shown for the areas directly inland of the extreme coast., i.e., the areas with highest precipitation.
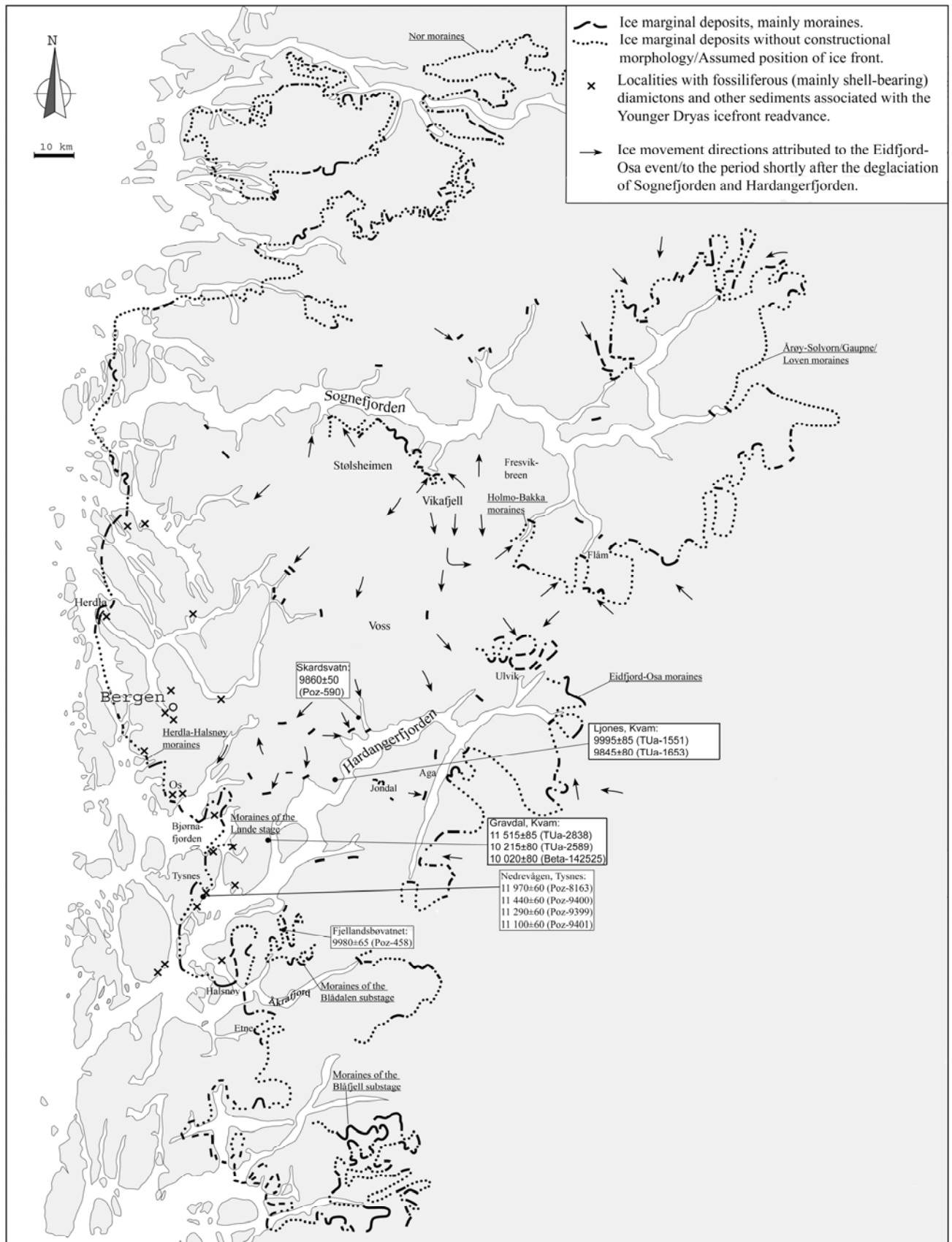
Fig. 2. The main moraine systems in western Norway. Based on Klovning (1963), Liestøl (1963), Mæland (1963), Undås (1963), Kvistad (1965), Holtedahl (1967, 1975), Skreden (1967), Anundsen & Simonsen (1968), Aarseth (1971), Anundsen (1972, 1985), Follestad (1972), Sindre (1973, 1974), Vorren (1973), Aarseth & Mangerud (1974), Bergstrøm (1975), Aa & Mangerud (1981), Aa (1982 a,b), Hamborg & Mangerud (1981), Hamborg (1983), Fareth (1987), Andersen et al. (1995), Sønstegaard et al. (1999), Mangerud (2000), Bakke et al. (2005). AMS dates undertaken on marine and terrestrial macrofossils along Hardangerfjorden, inside the Herdla-Halsnøy moraines, are shown (see Table 1).

related to the deglaciation history of the fjord.

However, imprints of an early post-deglaciation RSL rise of about 12 m or more are recorded at Vambheim, Bu and Hæreid, in the inner parts of the fjord (Paper 1), and possibly also at Ljones (Paper 3) and Gravdal (Fig. 11), in the mid part of the fjord (Fig. 1). This event interrupted the general, uplift-driven sea-level fall following immediately the deglaciation and is correlated with the YD transgression (see below). If correctly interpreted, this implies an early (Bølling/Allerød) deglaciation of the fjord. Furthermore, this implies that several of the marine limits along the fjord are associated with the maximum highstand of this sea-level rise, and that these marine limits probably are of late YD age (i.e., the same age as the majority of marine limits along the outer west coast). As a consequence, distinct changes in the shoreline gradients occur along the fjord, probably implying that several of the deep-seated basement faults along the fjord carry components of postglacial displacement (Paper 4).

The RSL event named the YD transgression has previously been documented through numerous investigations of sediment cores from isolation basins along the outer coastal areas of western and southwestern Norway (e.g., Anundsen, 1977, 1978; Thomsen, 1981; Anundsen and Fjeldskaar, 1983; Krzywinski and Stabell, 1984; Anundsen, 1985). The event was originally attributed to the Allerød period (e.g., Fægri, 1940, 1944), but is now generally considered to have started during the latter half of Allerød, or at the close of this period, and to have reached its maximum highstand late in the YD (e.g., Anundsen, 1977a, 1978, 1985; Lohne et al., 2004). The event is thus bracketed in age between Fairbank's (1989) meltwater pulse 1a and 1b (see e.g., Lohne et al. 2004). The vertical amplitude of the YD transgression typically varies between 10 and 13 m (e.g., Anundsen, 1985). Its eastern and western (geographical) limit in the region is not known (Anundsen, 1985). It is also unclear exactly which coastal areas outside western and southwestern Norway that experienced the event and which did not. In other parts of Norway, stratigraphical evidence of a RSL rise of between 6 and 10 m during the YD has been found in Finnmark (e.g., Leiknes, 2000). Minor sea-level oscillations during the YD have also been reported at the extreme northwestern coast (Svendsen and Mangerud, 1987). In the work of Anundsen and Fjeldskaar (1983), the YD transgression was associated with the YD ice-front readvance. Based on theoretical calculations, they interpreted this RSL event to be the result of the interplay between glacio-isostasy and geoidal eustasy (the latter signifying the gravitational attraction between the ocean and the growing ice masses), superimposed on the rising trend of glacio-eustatic sea level.

2. Objectives

The primary objective of this study has been to unravel, by means of stratigraphic methods, the early post-deglaciation RSL history of Hardangerfjorden. A key question that has been adressed is whether the sea-level changes were characterised by a monotonic fall (e.g., Hamborg 1983) or whether the initial RSL fall following immediately the deglaciation was interrupted by a RSL rise, similar to that found in the work of Helle et al. (1997). Other important aims have been:

- to obtain an absolute chronology for the deglaciation of the fjord,
- to reconstruct the highest post-deglaciation shorelines along the fjord and compare the shoreline delevelling pattern with that in adjacent fjords, and
- to detect potential zones of neotectonic faulting and to assess the role of neotectonism in the region's uplift history.
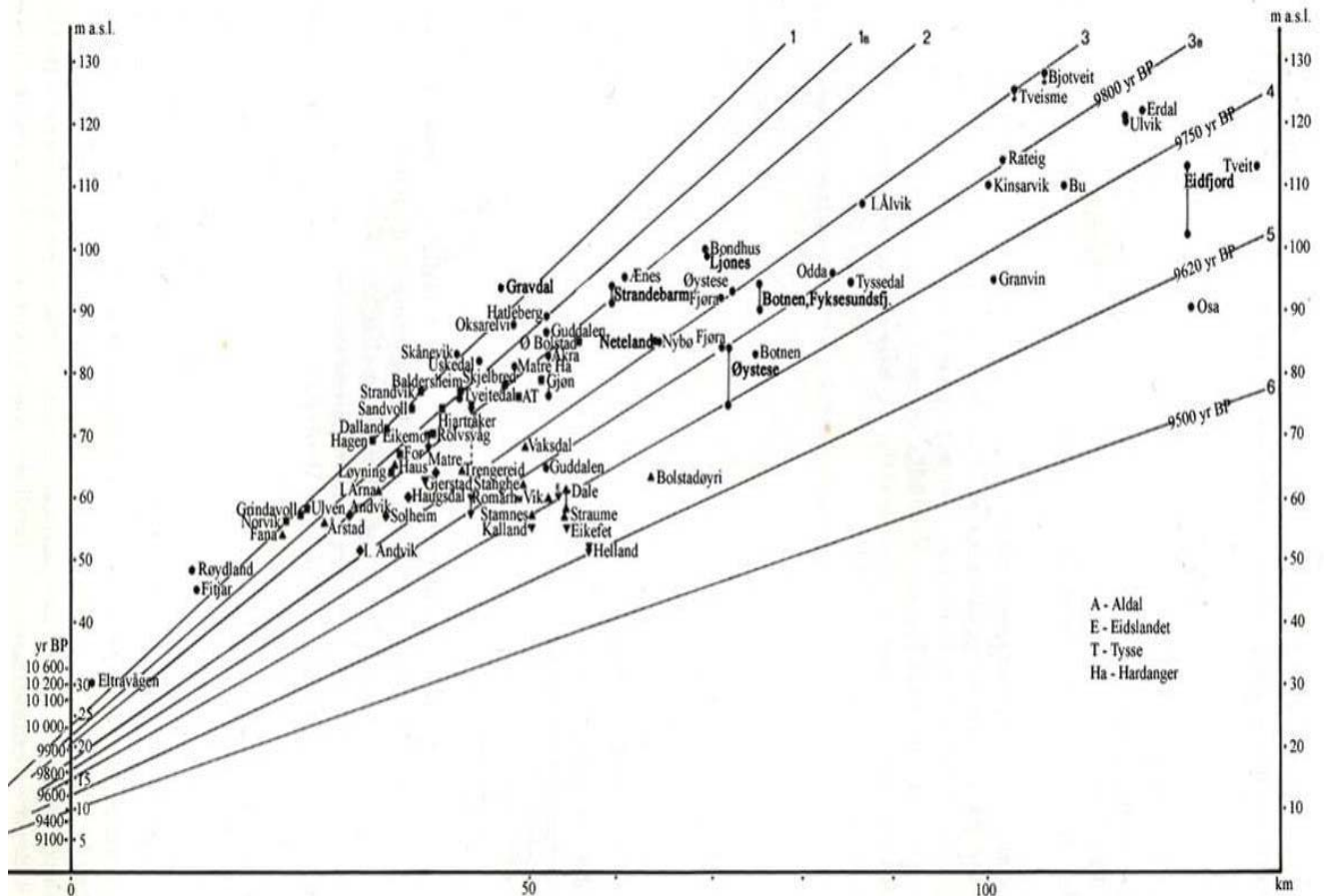
Fig. 3. Shoreline diagram for Hardangerfjorden (Hamborg, 1983). The shore levels are plotted on a projection plane that runs through Bømlo and which is striking N77°E. Also included in the diagram are the marine limits at Gravdal and Ljones as determined in the present study. Compare with the shoreline diagram (Fig. 6) in Paper 4. [Note that in Paper 4, the shore levels are plotted on a projection plane that runs through the southern tip of Sotra and which is striking N81°E].

3. Approach and methodology

In order to obtain the sea-level records in the present thesis, two distinctly different approaches were used; (1) litho- and biostratigraphic analyses of sediment cores from isolation basins and (2) sedimentological and georadar facies analysis of a marine moraine. The data in Paper 1 and 3 are provided by the former approach, whereas the data in Paper 2 are provided by the latter approach. In Paper 4, Late Weichselian sea-level data (mainly isolation basin data) from the region are correlated and combined into three shore-level profiles along SW Norway. The three profiles form, together with seismic reflection data and measurements on recent crustal movements, the basis for our detection of neotectonic fault zones in the region.

4. Papers

Paper 1:
Helle, S.K., Anundsen, K., Aasheim, S. & Haflidason, H. 1997: Indications of a Younger Dryas marine transgression in inner Hardanger, West Norway. *Norsk Geologisk Tidsskrift 77*, 101-117.

Paper 1 is based on my Cand. scient. thesis (Helle, 1993). In this work, the early post-deglaciation sea-level history at the head of Hardangerfjorden is reconstructed on the basis of litho- and biostratigraphic data from three isolation basins, situated 113, 119 and 128 m a.s.l. The total emergence since the deglaciation is estimated to about 128 m. The diatom stratigraphy in the two lowest basins exhibits distinct shifts interpreted as reflecting a RSL rise that interrupted the general,

uplift-driven sea-level fall. This contrasts the earlier, mainly geomorphologically-based reconstructions, where the early post-deglaciation RSL changes in the mid and inner parts of the fjord were thought as being characterised by a monotonic fall. The RSL rise is correlated with the YD transgression (Anundsen, 1985) implying, if correct, that Hardangerfjorden, or a major part of it, remained ice-free during the YD. This is in opposition to published reconstructions of the YD ice margin in the region. One implication of this is that the highest post-deglaciation shorelines at the head of the fjord probably developed during the maximum highstand of this RSL rise, rather than at the initial entry of the sea following immediately the deglaciation.

It is however important to note that the number of diatoms counted in the lower sediment interval exceeds generally 300 valves only in one of the basins, Vambheim-119. In the other basin, Bu-113, the number of diatoms counted in this interval is significantly less than 300. Hence, although important complementary information is provided by the diatom stratigraphy in Bu-113, the hypothesis of a marine transgression is largely based on the diatom stratigraphy in Vambheim-119. Moreover, the chronostratigraphic control is poor. The radiocarbon dates diverge significantly and give no clear evidence of the age of the sediments. In addition, the occurrence of Vedde Ash particles in the sediments is low and relatively indistinct, and they thus give no clear indications of the age of the sediments, either. Hence, the hypothesis of an ice free Hardangerfjorden during the YD heavily relies on a correct interpretation of the diatom and sediment stratigraphy in the investigated basins (see Section 6).

Looking beyond the discussion on the deglaciation and the earliest sea levels, the results presented in this paper add constrain on the timing of the onset of the early Holocene sea level regression and on the rapidity of the RSL fall.

Paper 2:
Helle, S.K. 2004: Sequence stratigraphy in a marine moraine at the head of Hardangerfjorden, western Norway: evidence for a high-frequency relative sea-level cycle. *Sedimentary Geology 164*, 251-282.

In this paper, the depositional architecture and stratigraphy of the Hæreid moraine in Eidfjord (Fig. 1) have been reconstructed in a high-resolution sequence stratigraphic framework. The aim of this investigation was to obtain a sea-level record independent of the isolation basin records, and thus, by using an approach totally different from that used by Helle et al. (1997), to test the hypothesis of an early post-deglaciation RSL rise at the head of the fjord. The analysis was based on data from ground-penetrating radar (GPR) profiles, outcrop sections and boreholes.

The paper describes in detail how the Hæreid moraine evolved under the control of a high-frequency RSL cycle, closely identified with the YD transgression. On the basis of the channel-fill/mouth-bar complex and the ravinement surface, which are two very important architectural elements of the moraine, the shoreline migration pattern can be reconstructed in great detail, both in a cross-sectional view and in a plan view. In the channel-fill/mouth-bar complex, the topset/foreset contacts can be seen as progressively climbing in a seaward direction until they change to become vertically aggrading at the distal part of the complex, shortly before they reach the level of the wave ravinement surface (GPR profiles C-C'and E-E'; Fig. 5 in Paper 2). The results from this study are thus completely in line with the diatom-inferred sea-level changes (Helle et al. 1997).

To the best of my knowledge, this is the first published paper where Quaternary deposits on *mainland* Fennoscandia have been subdivided using sequence stratigraphic principles. (Sequence stratigraphic methods have been used in studies of late Pleistocene and Holocene deposits in the southern Kattegat Sea and in the southwestern Baltic Sea; Jensen et al., 1997; Bennike et al., 2000; Jensen et al., 2002). The subdivision of the GPR profiles using sequence stratigraphic principles has the advantage that it helps predicting the occurrence of sedimentary facies and their distribution. This analytical approach gave the basis for the recognition of for example estuarine

facies in the GPR profiles (as well as in the outcrop section 1 and in the borehole section), which is a type of facies not commonly recognised in Quaternary deposits in Fennoscandia.

The example presented in this study may help to establish an architectural model with more general applications to other marine moraines (and other terraced deposits) that have undergone similar RSL changes. In addition to recording former ice-front positions, marine moraines provide excellent potential as high-resolution archives of past sea levels and shoreline migrations. It is my hope that this paper will stimulate not only to an increased focus on marine moraines but also to the use of sequence stratigraphic principles in the analysis of them.

Paper 3:
Helle, S.K., Rye, N., Stabell, B. & Haflidason, H.: Early postglacial relative sea-level changes in the mid Hardangerfjorden area, western Norway. Manuscript to be submitted to *Boreas*.

In this study, the hypothesis of a RSL rise along the fjord during the early post-deglaciation period is tested by investigating sediment cores from five isolation basins at Ljones, in the mid Hardangerfjorden area (Fig. 1), well inside the published reconstructions of the YD ice margin (Fig. 2). The pollen and diatom stratigraphy exhibit changes that may be suggestive of a major RSL cycle, similar to that found at the head of the fjord (Helle et al., 1997; Helle, 2004). However, the interpretation of the palaeoenvironment is complicated by the often low content of microfossils (and thus of sea level indicators) in the sediments. The isolation basin data therefore allow for alternative interpretations, implying that the basin water salinity may have been determined by other palaeoenvironmental factors, e.g., the amount of fresh water inflow into the fjord.

Another complicating factor is the poor chronological control on the basal sediments. Owing to the scarceness of terrestrial macrofossils in the basal sediments, bulk gyttja samples, aquatic plant material (spores and sea weed) and marine shells were mainly choosen as target material for ^{14}C dating. Since the basins (except for Øyjordsvatnet) are small and shallow, with rapid turnover and thorough ventilation, and are situated in a terrain with little or no carbonates present in the bedrock, the reservoir age of the aquatic plant material (see e.g., Olsson 1991) were considered to be small. However, parallel dates on bulk and macrofossil samples from the same sediment cores yielded ages that diverge significantly (the bulk dates generally showing younger ages than the macrofossil dates). Because of this inconsistency, little or no emphasis was put on the dates from the lower sediment core interval. In addition, the amount of Vedde Ash in the sediments is low, implying that the data remain rather inconclusive as regards the timing of the deglaciation and the earliest sea levels. This study clearly illustrates the difficulties with establishing an accurate chronology of early post-deglaciation sediments deposited in arctic-like conditions.

The early Holocene RSL history is fairly well constrained by the five sea level index points provided by the upper, dated isolation contacts in the five basins studied. The age of the isolation contact in the upper basin, Tveitatjørna (as inferred from ^{14}C dates, pollen data and a Vedde Ash microtephra horizon that may correlate to the Vedde Ash Bed) suggests that the onset of the early Holocene regression roughly coincides in time with the YD/Holocene transition. This implies that the maximum highstand shoreline in the study area and that at the outer coast, outside published reconstructions of the YD ice margin, is near-synchronous. The data from the other basins show that the RSL fall subsequent to the maximum highstand was very fast.

Paper 4:
Helle, S.K., Rye, N., Stabell, B., Prösch-Danielsen, L. & Hoel, C.: Neotectonic faulting and the Late Weichselian shoreline gradients in SW Norway. Manuscript submitted to *Journal of Geodynamics*.
In contrast to the three other papers, this paper does not focus on the deglaciation history of Hardangerfjorden. In this paper, we focus on the Late Weichselian shorelines in SW Norway and

on the role of neotectonism in the region`s uplift history. Sea-level data mainly derived from isolation basin studies are correlated and two shorelines are reconstructed along three profiles near Boknafjorden and Hardangerfjorden (Fig. 1, for location). The two shorelines are the Bølling/Allerød maximum lowstand shoreline, which formed prior to the YD transgression, and the YD maximum highstand shoreline, which formed at the end of this sea-level event. The majority of data fit into straight or slightly curvilinear, dipping trends. However, along all three profiles there are shore levels that sit as outliers, probably reflecting differential uplift rates on opposite sides of faults. This is consistent with observations previously reported as neotectonic `claims` in the region. Neotectonic faulting may be responsible for some of the deformational structures observed in seismic profiles of the fjord sediments, although the structures could also be explained in terms of other, non-tectonic processes. If our shoreline reconstructions (where we take into account faulting) are correct, this implies that the YD maximum highstand shoreline, on a regional scale, probably has a near constant gradient, close to c. 1.1 m/km. The implications of this for the modelling of glacio-isostatic adjustment processes, the YD peripheral ice-sheet profile and RSL changes in SW Norway are discussed.

5. The hypothesis of an ice free Hardangerfjorden during the Younger Dryas - discussed in the context of previous findings

In the present thesis, the stratigraphical evidence of a major RSL rise near the marine limit in Hardangerfjorden (Papers 1-3) has been taken to indicate that the fjord became deglaciated prior to the YD (i.e., in the Bølling/Allerød period) and that the fjord, or the main part of it, remained ice-free in the following period. The implications of this hypothesis for the reconstruction of shorelines and uplift pattern along the fjord have been discussed in Paper 4. However, in the papers, relatively little is said about the implications for the interpretation of the various glacial geological evidences in the area. I will therefore take the opportunity now to discuss this in more detail, focusing on earlier published observations that contradict our hypothesis. I will then, in the next section, critically evaluate our own interpretations of the sea level records along the fjord and discuss alternative interpretations that (possibly) accommodate better the existing YD ice model. Because of its relevance to the discussion later in the Introduction I will begin this section by giving a brief overview of the deglaciation history of the area prior to the YD.

5.1. The deglaciation history prior to the Younger Dryas

Investigations of marine sediment cores from the continental shelf off western Norway show that the Fennoscandian ice sheet had retreated from the central part of the Norwegian Channel by about 15 000 ^{14}C years BP (Lehman et al., 1991; Sejrup et al., 1994; 2000). The outermost coast of western Norway became ice free during the interval 14 000-12 600 ^{14}C years BP (Mangerud, 1977; Andersen, 1979; Anundsen, 1985; Paus, 1990; Houmark-Nielsen & Kjær, 2003). Based on stratigraphic evidence from Blomvåg (Fig. 1), Mangerud (1970, 1977, 1980) and Mangerud et al. (1979) postulated a short-lived ice-front readvance to a position seaward of the present-day coastline in the late Bølling/Older Dryas (Fig. 4). This interpretation was later disputed by Fjeldskaar et al. (1981), Krzywinski & Stabell (1984) and Anundsen (1985), mainly because ^{14}C dates from other sites nearby indicated that the outer coastal area had not been ice covered in the time interval given. However, a ^{14}C age plateaux centred around 12 600 ^{14}C years BP (Gulliksen et al., 1998) hinders a detailed chronostratigraphic correlation between the different sites (cf. Lohne, 2000). One cannot therefore, based on the ^{14}C dates alone, rule out the possibility of a late Bølling/Older Dryas ice-front readvance in the area. A minimum age for the final deglaciation of the outer coastal areas is provided by the isolation basin data from Os, where a firm sea level chronology recently has been established (Lohne et al., 2004) (Fig. 1, for location). Based on these
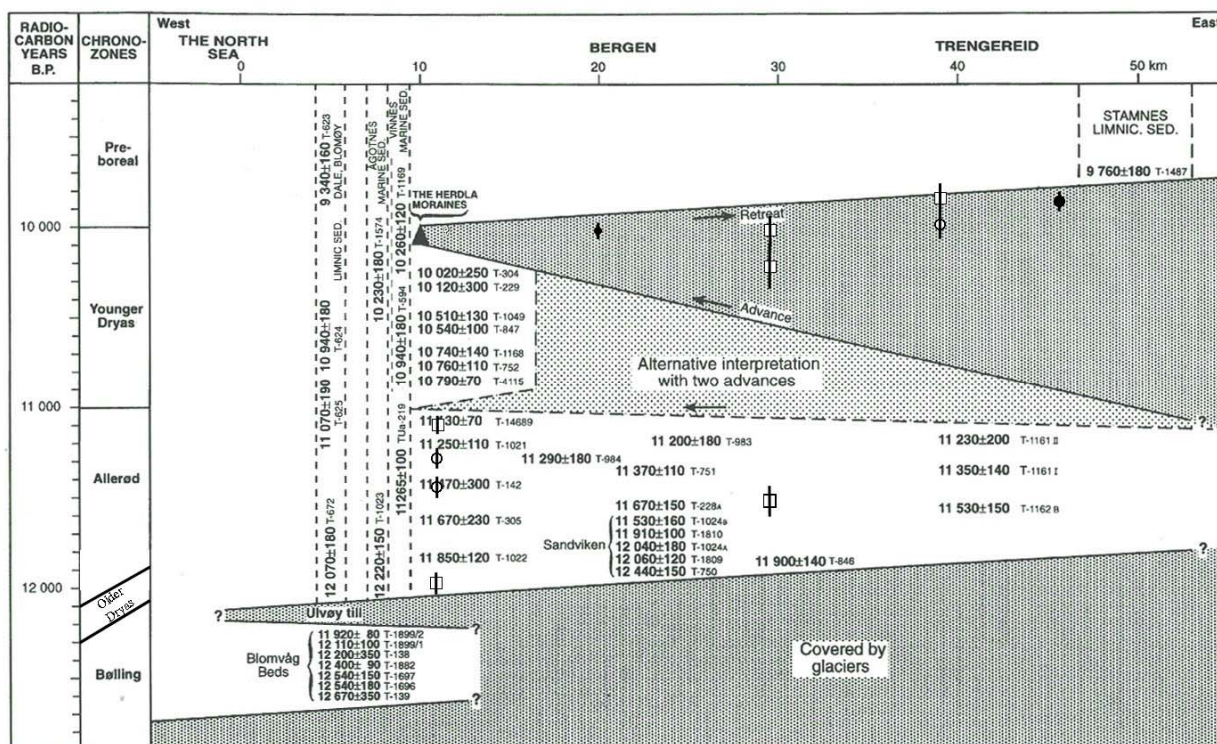
Fig. 4. Ice-front fluctuations in the Bergen district, according to Mangerud (2004). Also included are AMS dates on marine shells and terrestrial macrofossils from five sites proximal to the Herdla Moraines, apparently not overrun by the YD ice (J. Bakke, unpublished data; this study) (see Table 1 and Fig. 2). The five sites are (with increasing distance to the Herdla-Halsnøy moraines): Nedrevågen (Tysnes), Fjellandsbøvatnet (Uskedalen), Bersenmyra/Løkjen (Gravdal), Kolltveittjørna (Ljones) and Skardsvatnet (Fyksesund). Open dots – paired or whole shells. Open squares – fragmented shells. Filled dots – terrestrial macrofossils. Error bars: 1σ age range.

data, Lohne et al. (2004) estimated the timing of the deglaciation to around 14 450 calendar years BP (centred around ca. 12 500 ^{14}C years BP). A similar minimum age is provided by a ^{14}C date yielding 12 530±40 ^{14}C years BP (Beta-146547, corrected for a marine reservoir age of 380 years) from Fitjar, at the northwestern side of Stord (A. Nesje, pers.comm., 2000) (Fig. 1). The date is undertaken on marine shells (*Mya truncata*) in glaciomarine sediments interpreted as not being overrun by the ice.

In the period that followed, a rapid and extensive deglaciation took place (Fig. 4). Large areas on the landblock between Sognefjorden and Hardangerfjorden were probably ice free in the Allerød. Exactly where the front of the inland ice was located in this period is not known. Marine shells of Allerød age in sediments interpreted as being ice overrun during the YD are found at Trengereid, about 15 km east of Bergen (Mangerud, 1977) (Fig. 1). Since the mollusc fauna at this site includes low-arctic and boreal species, this indicates that they lived at some distance to the ice front (e.g., Mangerud, 1977; Andersen et al., 1995) (Fig. 4). Holtedahl (1975), Mangerud (1977), Sindre (1980), Hamborg & Mangerud (1981) and others indicated open fjords far inland in this period. Bouyant decoupling of the tidewater glaciers from the outer, shallow fjord thresholds, and subsequent iceberg calving, probably led to rapid, irreversible retreat of the ice front until the grounding line again shoaled. For Hardangerfjorden and Sognefjorden, with water depths of up to about 900 and 1300 m (Fig. 5), this would probably mean that only the shallowest tributary fjords were ice covered at the onset of the YD (unless the glacier had a positive mass balance sufficiently high to maintain its front in the deeper, central parts of the fjords during the Allerød, anchored to the narrow fjord sides).

Nevertheless, in order to refill Hardangerfjorden and Sognefjorden with ice during the YD, ice masses with a thickness of 1-2 km or more would probably had to accumulate in each fjord. This would require an immense production of ice on the adjacent mountain plateaus. In this respect

it is noteworthy that empirically-based palaeoclimate reconstructions (Bakke et al., 2005) and theoretical models (Lie et al., 2003) indicate that the ice production and the ice accumulation areas next to the fjords did *not* increase dramatically in this period.
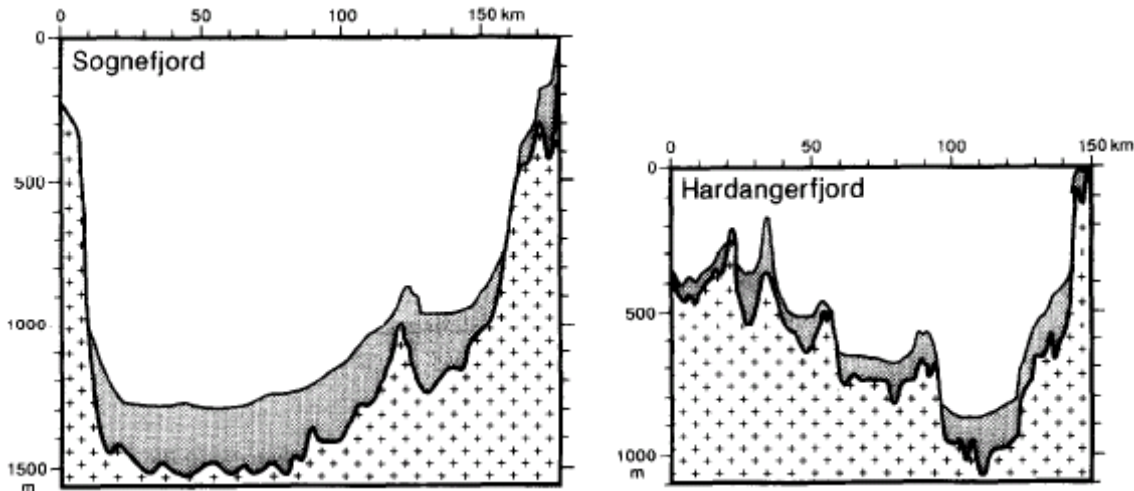
Fig. 5. Longitudinal profiles of the sediment distribution in Sognefjorden and Hardangerfjorden. Note that the vertical exaggeration is very large (from Aarseth, 1997).

5.2. *Published reconstructions of the Younger Dryas ice margin*

Across Hardangerfjorden and northwards to Sognefjorden, the moraines considered as the YD moraines are referred to as the Herdla-Halsnøy moraines (e.g., Mangerud 2000) (Fig. 2). The most prominent moraines are located at the outer parts of Herdlefjorden (the Herdla moraine) and Hardangerfjorden (the Halsnøy moraine), and at Os (the Ulven moraine) (Aarseth & Mangerud, 1974; Holtedahl, 1975; Andersen et al., 1995; Aarseth et al., 1997) (Fig. 2). Generally, however, the moraines are small, typically 1-5 m high, and discontinuous, and they usually consist of one ridge, although in rare cases they occur several in parallel (Aarseth & Mangerud, 1974; Andersen et al., 1995).

In Bjørnafjorden (Fig. 1), the YD ice margin is reconstructed from the projection of moraines at Strandvik and at Tysnes (Aarseth & Mangerud, 1974). On the peninsula between Hardangerfjorden and the head of Bjørnafjorden (Fig. 1) there is a predominance of westerly directed glacial striations. Aarseth & Mangerud (1974), Mangerud (2000) and others attributed most or all of these to the YD ice-front readvance, implying that the peninsula (including a
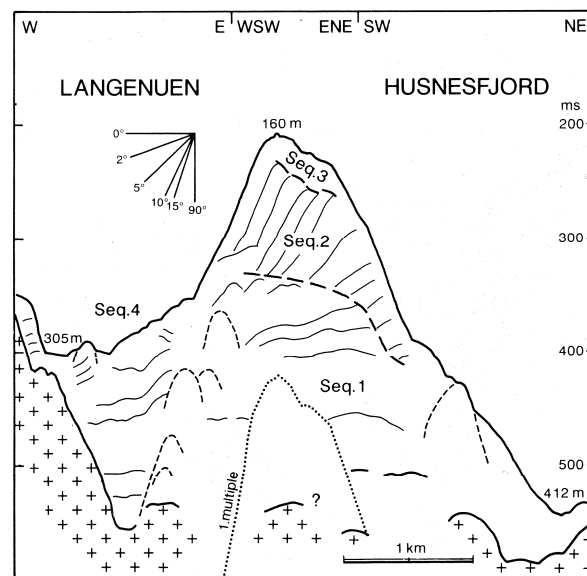
Fig. 6. Interpretation of seismic profile 84-6 between the islands Halsnøy and Huglo. For location, see Fig. 7. After Hoel (1992) and Aarseth et al. (1997).
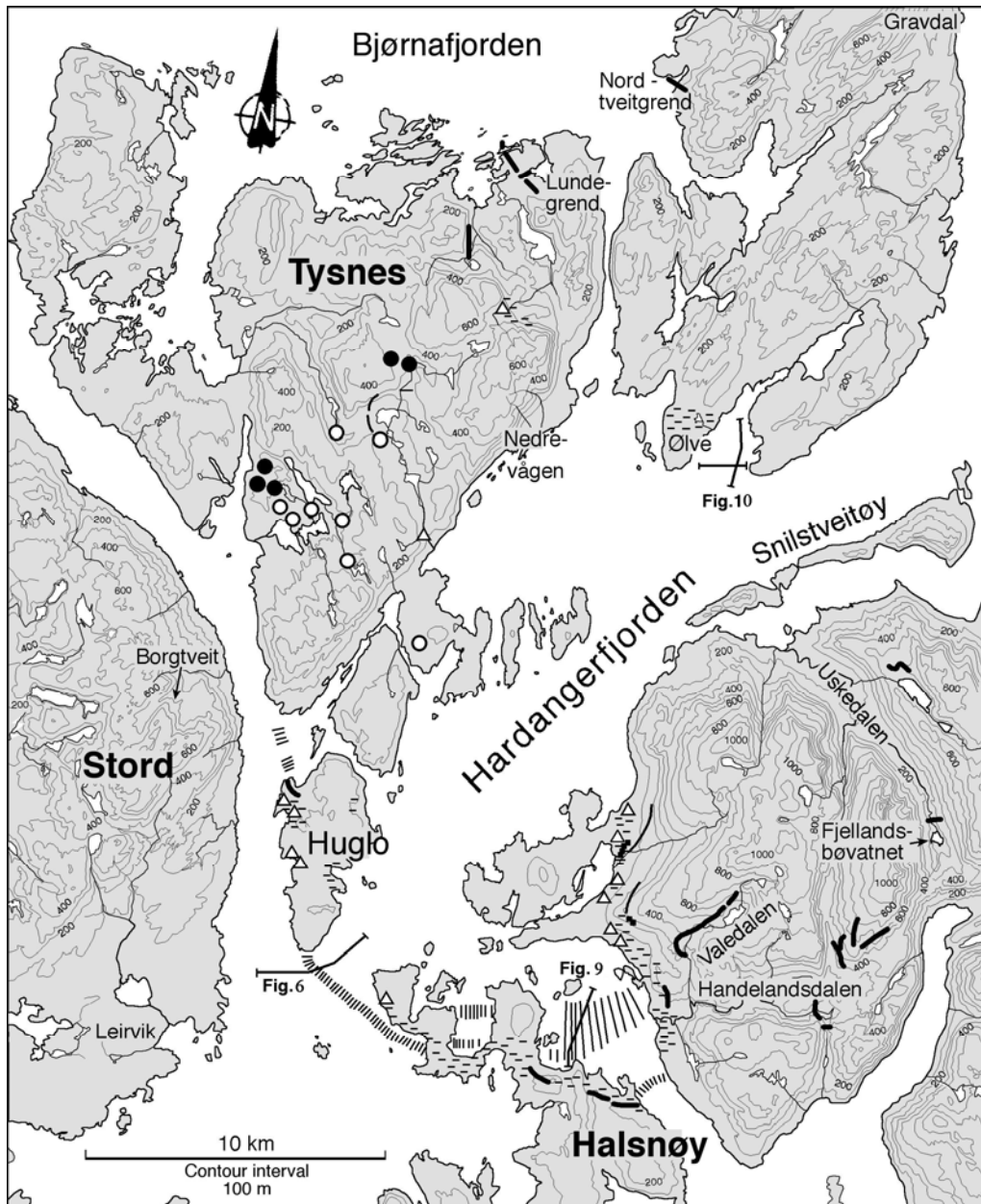
Fig. 7. Marginal moraines and other morainal accumulations in the outer Hardangerfjorden area (for location, see Fig. 1). Bold line: moraine ridge above present sea level. Hatched line: submarine moraine ridge/marginal morainal complex. Dashes and triangles: boulder clay/marginal morainal complex above present sea-level (see text). The dots on Tysnes refer to the basins investigated by Lohne et al. (2003) and Lohne (2006). Filled dots: basins where Lateglacial deposits, including Vedde Ash and organic-enriched Allerød sediments, are found. Open dots: basins where only deposits postdating the Vedde Ash fall are found.

mountainous area with altitudes of 5-600 m and more) was crossed by an ice coming from Hardangerfjorden in this period. In the central part of Tysnes, distinct moraines with the ice-proximal side facing Hardangerfjorden have been attributed to the same ice-front readvance (Undås, 1963) (Figs. 2 and 7).

The Halsnøy moraine (Fig. 7) is located on a bedrock threshold and reflects a grounded ice front (Holtedahl, 1967). Seismic reflection studies between the islands Huglo and Halsnøy have revealed sediment thicknesses of up to 240 m and show that the submarine ridge here contains two, vertically stacked foresets separated by an angular unconformity and with depositional dips directed (west)southwestwards (Hoel, 1992; Aarseth et al., 1997) (Fig. 6). The western part of the island Halsnøy is to a great extent covered by a clayey diamicton, interpreted as till (Holtedahl, 1975). In the southeastern part, well-defined moraine ridges occur (Fig. 7). On the mountain plateau on the mainland northeast of the Halsnøy island (Fig. 7), an ice cap existed during the YD,

feeding a system of outlet glaciers directed radially outwards from this centre (Follestad, 1972) (Fig. 8). According to this reconstruction, all outlet glaciers, except for those in the south, drained into the adjacent fjord glaciers. The extent of the outlet glaciers in the south is delimited down-valley by a set of low-lying moraines, interpreted to be lateral moraines of an eastern branch of the Hardangerfjorden glacier. This interpretation is based on the observations that a shell-bearing clayey diamicton interpreted as till occurs on the southwest side of these moraines and that the youngest glacial striations here have a southerly alignment (Follestad, 1972).
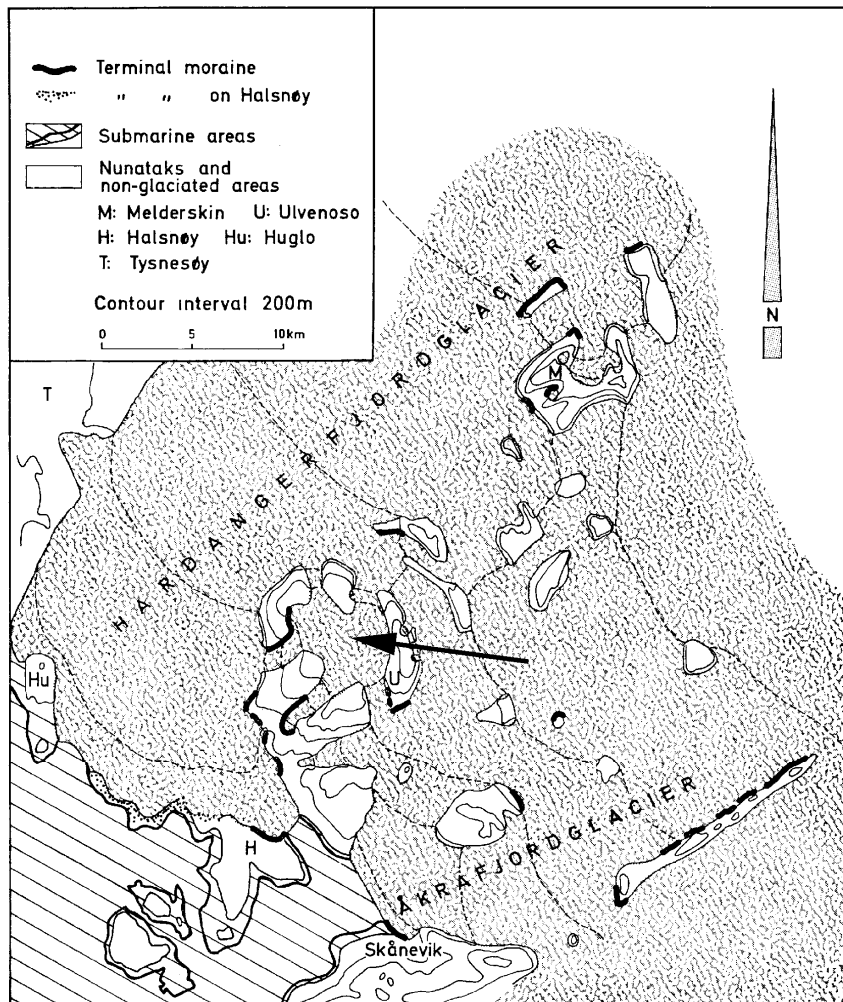
Figure 8. The YD ice margin in the Halsnøy area, according to Follestad (1972). Arrow marks the ice accumulation centre discussed in the text.

        On the proximal side of the Herdla-Halsnøy moraines, the ice-front readvance is age determined by means of ^{14}C dates on shells in diamictons interpreted as till, or in other sediments interpreted as ice overrun (shown as crosses in Fig. 2). A YD age of the moraines around the outer Hardangerfjorden area is inferred from ^{14}C shell dates from the head of Bjørnafjorden, from the eastside of Tysnes, from Ølve and from Valen (Aarseth & Mangerud, 1974; Holtedahl, 1975; Mangerud, 2000) (Figs. 2 and 7). The youngest date is from the eastside of Tysnes, yielding 9940±160 ^{14}C years BP (Holtedahl, 1975). In addition, a number of ^{14}C dates from the basal sediments in isolation basins proximal to the moraines, yielding ages close to the YD/Preboreal transition, has been taken to indicate an early Preboreal deglaciation of the fjord (e.g., Romundset, 2004).

        On the distal side of the moraines, glaciomarine silt interpreted as meltwater deposits has been dated at Os (Bondevik & Mangerud, 2002), at Vinnes (Øvstedahl & Aarseth, 1975; Aarseth et al., 1997), at Tysnes (Lohne et al., 2003) and at Halsnøy (Lohne, 2006) (see Fig. 2, for location).

Based on these investigations it is concluded that the readvance ended at the YD/Preboreal transition (Mangerud, 2000; Bondevik & Mangerud, 2002; Mangerud, 2004) (Fig. 4). An important test of previous reconstructions of the ice margin on Tysnes has recently been made by Lohne et al. (2003) and Lohne (2006). They show that isolation basins and other basins on the northwestern part of the island (i.e., basins located farthest away from Hardangerfjorden) contain Lateglacial deposits (including organic-enriched Allerød sediments and the Vedde Ash Bed), and that basins located on the central and southeastern parts only contain deposits that postdate the Vedde Ash fall (Fig. 7). In some of the distal basins, thick silt accumulations of YD age are identified, interpreted as meltwater deposits from the glacier (Lohne et al., 2003). These results, together with the other observations from the area, led Lohne et al. (2003) to conclude that the central and southeastern parts were covered by the Hardangerfjorden glacier during the late YD, consistent with the conventional ice model. Compared to previous reconstructions, however, their reconstruction implies a slightly more extensive ice cover on the southern part of the island during the YD glaciation maximum (with the ice front located between the filled and open dots in Fig. 7, i.e., partly beyond the moraines of Undås, 1963).

The envelope of ^{14}C dates in Fig. 4 also allows for a fast first readvance in the early YD (Andersen et al., 1995). In this respect it should be mentioned that shell-bearing diamictons associated with the YD icefront readvance have been found also distal to the Herdla-Halsnøy moraines. One km distal to the moraine at Strandvik, in Bjørnafjorden (Figs. 1 and 2), crushed barnacles of *Balanus balanus* in a till-like diamicton are dated to 10 840±190 years BP (Aarseth et al., 1997). Aarseth et al. (1997) suggested that this diamicton may either be a till deposited by the inland ice (implying that the glacier advanced beyond the moraine during the YD) or it may be a (glaciomarine) sediment disturbed by iceberg ploughing. Near Leirvik, at the southeastern coast of Stord, about 10 km distal to the Halsnøy moraine (Fig. 7), shells in sediments interpreted as till are ^{14}C dated to between 11 000 and 12 000 years BP (Sindre, 1980). The coastal areas of Stord were, according to Sindre, overrun by the inland ice during the early YD. Mangerud (2000) indicated, on the other hand, that the fossiliferous diamict sediments of Sindre (1980) may instead be of glaciomarine origin possibly disturbed by icebergs and that the YD ice front may thus not have reached Stord. On the same island, at Borgtveit (Fig. 7), a ^{14}C date on marine shells from glaciomarine sediments apparently undisturbed by the ice, yielding 10 600±140 years BP (or 10 430±140 years BP if corrected for a marine reservoir age of 610 years; cf. Bondevik et al., 1999), indicate that the ice front did not reach the island during the late YD (Genes, 1978; Sindre, 1980). From seismic profiles, there are no indications of an early YD ice advance beyond the Halsnøy moraine (Aarseth et al., 1997).

5.3. Discussion of implications of our hypothesis for the deposits in the outer Hardangerfjorden area
North of Hardangerfjorden, the YD ice sheet was nourished mainly from ice-accumulation centres on the landblock between Hardangerfjorden and Sognefjorden, some of which were located relatively close to the moraines. Ice accumulation centres in the Gullfjellet area (cf. Skår, 1975) and in the Botnavatn area (cf. Hamborg & Mangerud, 1981) (Fig. 1) probably nourished the branch of the ice sheet responsible for the moraines in the Os-Fusafjorden area (including the Ulven moraine at Os; Fig. 2) (Bakke et al., 2005). Also south of Hardangerfjorden, there were ice accumulation centres located directly proximal to the YD moraines, e.g., on the Folgefonna peninsula and in the mountains south of Åkrafjorden (Fig. 1). An ice accumulation centre in the latter area probably nourished the outlet glaciers responsible for the moraines in Etne (cf. Anundsen, 1977b) (Fig. 2). These moraines and the moraines in the Os-Fusafjorden area may thus have formed independent of whether there was ice in Hardangerfjorden or not during the YD. The conflict that arises from our hypothesis of an ice free Hardangerfjorden with respect to the published reconstructions of the YD ice margin pertains mainly to the correlation of moraines in the Bjørnafjorden area and southwards

across Hardangerfjorden. If our hypothesis is correct, then the moraines here must pre-date the YD and/or have formed by land-based/local glaciers (cf. Bakke et al., 2005).

According to the computations of Lie et al. (2003) (cf. Fig. 5 in their work), the mountain plateau northeast of Halsnøy (Fig. 7) has a relatively high sensitivity for glacierization. The plateau has probably undergone a complex glaciation history, covered with ice caps both in periods with and in periods without glaciers in the fjords. Observations of glacial striations and of proglacial and ice-contact sediments on Snilstveitøy and adjacent areas north of this mountain plateau suggest that outlet glaciers originating from the plateau still remained in the tributary valleys after the deglaciation of the main fjord (Follestad, 1972). Originally, the moraine ridges on the southeastern part of Halsnøy were interpreted as deposited by outlet glaciers or local valley glaciers descending from the same mountain plateau (Rekstad, 1906). Undås (1963), however, suggested that these moraines were deposited by an eastern branch of the Hardangerfjorden glacier during the YD, agreeably to the conclusions of Holtedahl, (1967) and Follestad (1972), but that valley glaciers still remained, terminating in the sound between the island and the mainland after the fjord glacier had melted back. The latter is not in congruence with the conclusions of Holtedahl (1967) and Follestad (1972), who indicated that the sound was last occupied by a glacier coming from Hardangerfjorden.

On the mainland, directly west of Valedalen and Handelandsdalen (Fig. 7), near the present shoreline, there is a few (older) glacial striations directed towards west and northwest (i.e., towards Hardangerfjorden) (cf. Plate 1 in Follestad, 1972). Most likely, these were formed in a period when the sound was not occupied by the Hardangerfjorden glacier, by valley glaciers or piedmont-like glaciers extending into the sound between Halsnøy and the mainland. In the sound, a distinct (25-30 m high) sedimentary ridge occurs, together with a series of minor ridges in the south (Hoel, 1992) (Fig. 9). The position of these ridges, outside the mouth of the valley Valedalen (Fig. 7), may not be coincidental and could fit with the idea that these ridges (or at least the large ridge in the north) were deposited by a dynamically active (and advancing?) valley glacier. Aarseth et al. (1997), however, interpreted these ridges as recessional moraines, possibly DeGeer moraines, deposited by the Hardangerfjorden glacier, after the eastern branch had retreated from its maximum YD position. Their interpretation accommodates better the observations from the mainland area suggesting that the sound was last occupied by an ice coming from Hardangerfjorden (cf. Follestad,
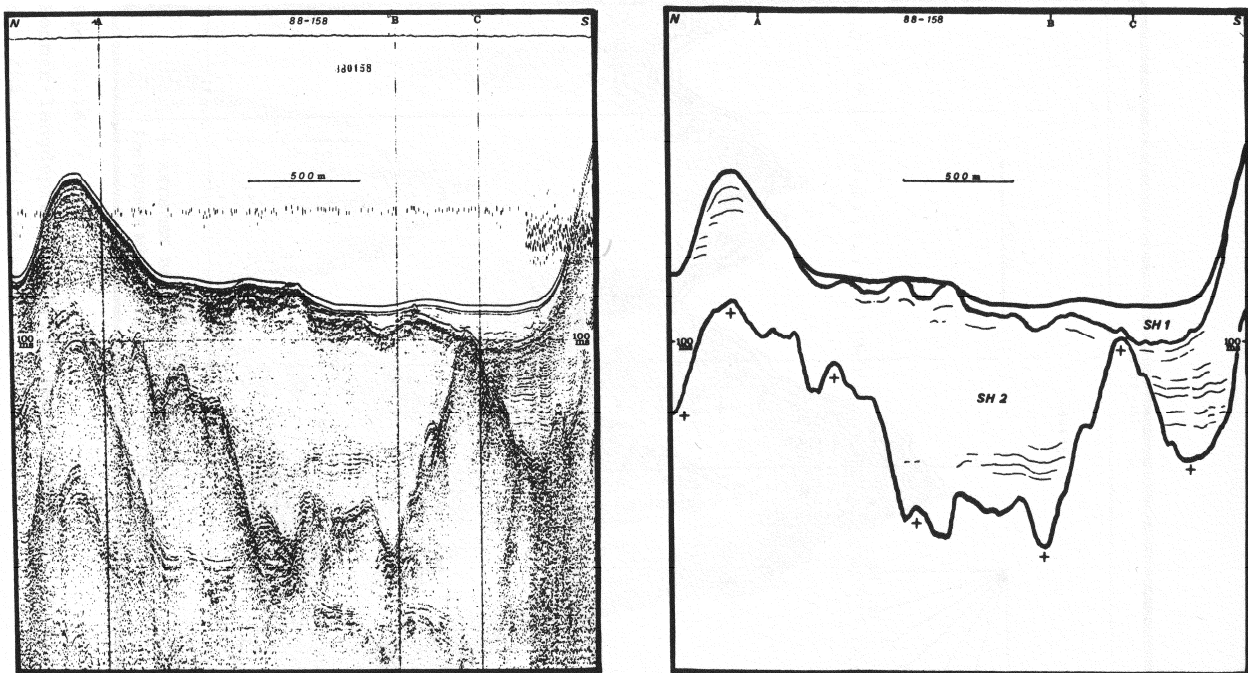

Fig. 9. Seismic profile 88-158 in the sound between Halsnøy and Valedalen. The interpretation to the right (Hoel, 1992). + + + marks acoustic basement (assumed bedrock). ms: milliseconds two-way-travel time. Max. sediment thickness: ca. 50-55 m. For location, see Fig. 7.

1972). Their interpretation also accommodates better the observations from the other parts of the outer Hardangerfjorden area, including Lohne et al.`s (2003) data from Tysnes. Nevertheless, the interpretations that outlet glaciers or local valley glaciers extended into (Undås, 1963) and crossed (Rekstad, 1906) the sound are interesting in view of our hypothesis of an ice free Hardangerfjorden during the YD. In particular, Rekstad`s (1906) interpretation is interesting because this opens for the possibility that the Halsnøy moraine is of polyepisodic origin and includes not only moraines that were formed by a glacier in the main fjord but also moraines built by valley glaciers or piedmont-like glaciers that extended into and crossed the sound after the main fjord was deglaciated. However, as already noted, observations on the mainland (cf. Follestad, 1972) do not support this interpretation. A detailed reconstruction of the stratigraphic and sedimentary architecture of the Halsnøy moraine and adjacent deposits may contribute to solve some of the puzzles associated with the deglaciation history of Hardangerfjorden.

Northeast of the aforementioned mountain plateau, there is a moraine ridge system representing a southwestern branch of the Folgefonna glacier (Fig. 2) and which is attributed to an early Preboreal ice-front readvance (Follestad, 1972). This event, termed the Blådalen substage, was tentatively correlated with the Eidfjord-Osa event (Follestad, 1972) (see below). A minimum age of this moraine system is recently provided by an AMS date, yielding 9980±65 ^{14}C years BP (11250-11750 cal. years BP), on a terrestrial macrofossil from a lake basin (Fjellandsbøvatnet, Fig. 7) proximally to the moraines (J. Bakke, unpublished data). This suggests that the moraines belonging to the Blådalen substage have a closely similar age as the Herdla-Halsnøy moraines (and as several of the Eidfjord-Osa moraines, see below).

On the opposite (northern) side of Hardangerfjorden, several distinct marginal moraines are observed directly proximal to the Herdla-Halsnøy moraines (Aarseth, 1971). At Lundegrend, on the northeastern side of Tysnes (Fig. 1), distinct ridges, up to three in parallel, can be traced for about 1.5 km. A distinct (150-200 m long) moraine ridge is also identified at Nordtveitgrend, at the head of Bjørnafjorden (Aarseth, 1971) (Fig. 7). Two ^{14}C dates on marine shells from diamictons interpreted as till on the proximal side of these moraines yielded ages of respectively 10 570±100 and 11 320±180 years BP (Aarseth, 1971; Aarseth & Mangerud, 1974). Aarseth (1971) interpreted the moraines (which all lie below 60-70 m a.s.l.) as being deposited on each side of a marginal tongue of the Hardangerfjorden glacier (the Lunde stage) shortly after the glacier had retreated from the Halsnøy moraine (i.e., during the early Preboreal).

At the inlet of the bay east of Ølve, there is a well-defined sedimentary ridge buried beneath thick (glacio-) marine sediments (Fig. 10). The ridge is mound-shaped, a few hundred meters in diameter and up to 15 m in height (Fig. 10). The seismic profiles reveal little information about the internal stratigraphy of the ridge, but the upper boundary is thought to be an erosional surface (Hoel, 1992). Onshore, at Ølve (Fig. 1), directly northwest of the ridge, a belt of compacted, faulted and folded marine sediments extends across the promontory (Aarseth, 1971; Aarseth & Mangerud, 1974). Shell fragments collected from these sediments are ^{14}C dated to 11 230±180 years BP and the sediments are interpreted as being glaciotectonized during the YD ice readvance (Aarseth, 1971; Aarseth & Mangerud, 1974). The buried ridge is attributed to the same readvance and is interpreted as being formed by subglacial erosion and deformation (Hoel, 1992).

However, the possibility that this is a terminal moraine formed by a glacier that stabilized at the bay inlet after the ice had retreated from the main fjord channel should also be taken into consideration. Since the fjord directly outside the bay is very deep (in some places nearly 700 m), this may have led to a faster ice-front retreat here (due to calving) than on the mainland, implying that ice masses may have been left isolated on the peninsula after the main fjord channel became ice free. Also the other, low-lying moraines in the north may have been deposited by isolated ice
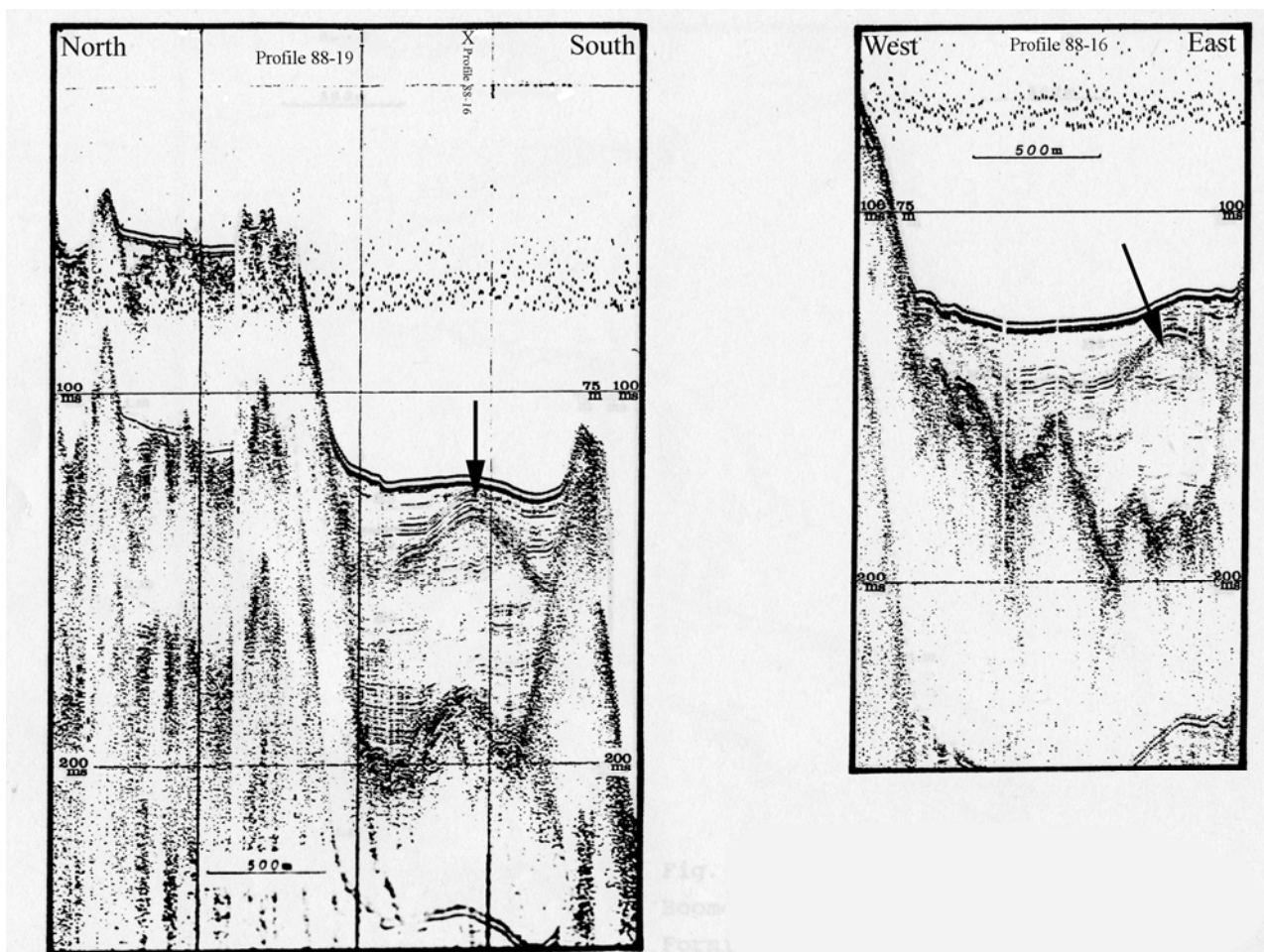
Fig. 10. Seismic profiles from the bay outside Ølve (Hoel, 1992). For location, see Fig. 7. Arrows mark the sedimentary ridge buried below thick glaciomarine sediments (see text).

masses that later became re-activated. The interpretation that the moraines of the Lunde stage were formed by an ice tongue that was dynamically connected to the Hardangerfjorden glacier is mainly based on a general assessment of the topography and not on observations of glacial striations (cf. Aarseth, 1971, p. 50). The glacier(s) that were responsible for the moraines must have been very thin, at least in the marginal parts, and with a low-graded (piedmont-like?) ice profile. The shell-bearing diamictons proximal to the moraines at Lundegrend and at Nordtveitgrend, as well as the deformed sediments at Ølve could thus reflect the work of local ice masses; a scenario that conform better with our idea of an ice free Hardangerfjorden during the YD. On the other hand, it is not easy to see how local ice masses could have been sustained in this lowland area, far away from any known ice-accumulation centre (unless ice caps in the Botnavatnet area and on the Folgefonna peninsula caused the precipitation maximum (Fig. 1) to migrate westward, causing in turn local ice masses to build up in the outer coastal areas, and leaving the interior in a precipitation shadow, see below). This may support the hypothesis of Aarseth (1971) of a connection to a fjord glacier.

Most problematic is to explain the data from Tysnes in the context of an ice free fjord during the YD. Lohne et al.`s (2003) and Lohne`s (2006) observations, where Lateglacial sediments (including the Vedde Ash Bed), despite being thoroughly searched for, are found only on the northwestern part of the island (Fig. 7), represent an important, independent test of previous reconstructions. Certainly, the possibility that the stratigraphic level where Lateglacial sediments occur was not reached during coring of the basins nearest Hardangerfjorden cannot be completely ruled out. There is also (at least in theory) a possibility that the distribution of Lateglacial sediments on the island is unrelated to the position of the former ice margin, i.e., that the lack of Vedde Ash and organic-enriched Allerød sediments in the basins nearest Hardangerfjorden is not a result of an

ice cover during the YD (given the fact that there are basins also distal to the Herdla-Halsnøy moraines where only Holocene deposits are found). However, the observations are simplest explained in terms of an ice-filled fjord during the YD (cf. Lohne et al., 2003) and the abovementioned explanations must be considered only as *ad hoc* solutions to accommodate our hypothesis.

Also the observation of thick silt accumulations in some of the distal basins of Lohne et al. (2003), dating from the late YD and interpreted as meltwater deposits analogous to the Grindavoll Silt (cf. Bondevik & Mangerud, 2002), may add support to the notion that these basins were located in front of the ice margin. It is well known that, under appropriate conditions, the occurrence of silt in proglacial lake deposits can be used as a signature for the existence of glaciers within the catchment (e.g., Karlèn, 1976; 1981; Leonard, 1985; Dahl et al., 2003). In Holocene lake deposits, a change from a situation with glaciers in the catchment to a situation without glaciers is typically seen as a shift from bluish-grey sandy and/or clayey silt to gyttja (Dahl et al., 2003). However, moderate differences in geomorphology and process activity in the lakeshore region around alpine/proglacial lakes can also significantly affect the lake sediment composition (Rubensdotter & Rosqvist, 2003). This implies that non-glacial processes around proglacial lakes deposit minerogenic sediment layers with similar characteristics (high density, low organic content) as layers interpreted as having a glaciofluvial origin, thus obscuring the potential glacial sediment signal (Rubensdotter & Rosqvist, 2003).

The situation is even more complex for basins located below the marine limit, where the basin deposits in addition also record sea-level changes. Among the basins studied by Bondevik & Mangerud (2002), Lohne et al. (2003) and Lohne et al. (2004), it is in those located below the marine limit that the thickest silt accumulations in general are found, representing marine transgressive sediments of late YD age. As discussed in the following, the high rates of silt accumulation in the basins below the marine limit may, at least to some extent, be related to the YD transgression, implying that there may not be a direct link between the sediment production in the upstream reaches and the high rates of silt accumulation in the basins.

Siltation is a well-known phenomenon along modern transgressive coasts (estuaries), including in microtidal environments, where it frequently leads to reduced vessel access, requiring dredging of channels to maintain safe navigation. There is a large amount of literature describing estuarine siltation processes in detail. [For some recent works, it is referred to e.g., Winterwerp (2002), Allen (2004), Ganju et al. (2004), Pekar et al. (2004), Pontee et al. (2004), Prandle (2004) and Prandle et al. (2005)]. Estuaries typically accumulate sediments rapidly. Significant estuarine siltation most likely occurs via the entrainment of fine marine sediments (Prandle, 2004), where the fine sediments are trapped by tidal pumping. In extreme cases, deposition rates of up to 30 cm/year of fine-grained sediments (mostly silts and clays) have been recorded in microtidal, non-glacial environments (Pekar et al., 2004). However, large variations in sediment mass accumulation occur among sites (Colman et al., 2002). High freshwater discharge will often lead to reduced siltation rates since this can flush sediments out of the estuarine system (e.g., Pontee et al., 2004). Since siltation often increases in a positive feedback fashion (tidal pumping is often increasing as the system is silting, e.g., Wolanski et al., 2001) this may explain why the significant siltation in some of the basins investigated by Bondevik & Mangerud (2002), Lohne et al. (2003) and Lohne et al. (2004) started first at a late stage of the YD transgression, well after marine sediments began to accumulate in the basins. The 2-m-thick laminated silty clay observed in the near subsurface of the Hæreid moraine, interpreted as a highstand mud drape (Paper 2), may have a parallel genesis to the Grindavoll Silt (and may even represent an isochronous facies?). It is also worth noting that at Vinnes (Fig. 1, for location) there is a 27-m-thick succession of marine and glacimarine sediments, mainly silt and clay, deposited in the late Allerød-late Holocene time interval (Øvstedahl & Aarseth, 1975; Aarseth et al., 1997; Mangerud, 2000). Here, the siltation rates were very high also during the late Allerød and throughout much of the Holocene, although not as high as during the

YD (Øvstedahl & Aarseth, 1975; Aarseth et al., 1997). It is also interesting to note that the silt of YD age interpreted as meltwater deposits is rich in terrestrial plant macrofossils. The surface of the deposit, ca. 4 m a.s.l., emerged from the sea only a few thousand years ago and the deposit probably includes sediments originating from the YD transgression, as well as from the mid-Holocene (Tapes) transgression. Hence, for the basins located below the marine limit, the contribution from estuarine siltation processes should be known before firm conclusions on meltwater input from the upland catchments are drawn.

Estuarine siltation processes cannot, however, explain the relatively thick accumulations of silt in the basins located above the marine limit and within the inferred drainage area of the Hardangerfjorden glacier meltwater (Lohne et al., 2003). The fact that relatively thick silt accumulations also occur above the marine limit (albeit less thick than in the basins below) shows that the upland catchment after all contributed significant amounts of silt to the basins. Hence, unless the silt can be explained in terms of non-glacial processes (cf. Rubensdotter & Rosqvist, 2003), or reflects meltwater input from local glaciers, or reflects a climatically-induced reduction of the organic production, the occurrence of silt provides evidence for a late YD ice sheet maximum in the area, in agreement with the interpretation of Lohne et al. (2003). Altogether, the observations from Tysnes are difficult to reconcile with the hypothesis of an open, ice free fjord. One possible solution to this conflict is that the RSL records that form the basis for our hypothesis is erroneously interpreted and that the fjord was filled with glacier ice during the YD (see Section 6).

I also find it appropriate to provide some comments on the interpretation of the radiocarbon-dated shell-bearing sediments (mainly diamictons) along the outer part of the fjord. So far, the diamictic sediments of Allerød and YD age on the proximal side are invariably interpreted as till, deposited by the YD Scandinavian ice sheet (e.g., Aarseth & Mangerud, 1974; Holtedahl, 1975; Andersen et al., 1995; Mangerud, 2000), whereas those on the distal side have been interpreted either as till or as glaciomarine diamictons (Sindre, 1980; Aarseth et al., 1997; Mangerud, 2000). It has been argued that the diamictons at the proximal side of the moraines are more compact than those on the distal side (Mangerud, 2000). At one site, Ølve (Fig. 7), oedometer tests clearly show that the clayey sediments here are very compact (Aarseth, 1971). However, systematic geotechnical investigations of the sediments in the outer Hardangerfjorden area are lacking.

Often it may be difficult to discern sediments compacted under the load of a glacier from sediments compacted by other processes. Given the fact that all the ^{14}C dated diamictons associated with the YD ice advance in the Hardangerfjorden area are located at or below the local marine limit, processes operating in near-shore, glaciomarine environments should also be taken into consideration. An extensive and pervasive sea-ice cover probably existed in the fjord after the deglaciation [as can be inferred from e.g., the sea ice indicators in the diatom records in Paper 3 and from the observation of boulder barricades along the fjord sides (e.g., Helle, 1993, p. 9)]. The shores were probably exposed to the action of floating ice, as well as to intense frost shattering and collapse from the steep fjord sides. In addition, icebergs that calved in the fjord probably stranded at the outer, shallow threshold areas. In periods with substantial calving the narrow fjord inlet may have been completely blocked by icebergs. In these environments it is likely that compact, till-like diamictons and other lithofacies that easily could be mistaken as being overrun by the inland ice were deposited. The load of the overburden sediments will also lead to compaction. For example, the deposits below the Grindavoll Silt are compacted and difficult to penetrate with cores (Bondevik & Mangerud, 2002). Non-mechanical processes (e.g., drying) should also be taken into consideration. As a consequence of uplift of an area and lowering of the ground-water level, the loss of water may have caused the sediments to harden and, in some cases (depending on the mineralogic composition of the finest fractions), to have cemented into a concrete-like mass (see e.g., Nenonen et al., 2000).

In a 200-m-wide and 4-m-high outcrop section on the eastside of Tysnes (near the site described by Mangerud, 2000) (see Fig. 7, for location), J. Bakke (pers. comm., 2006) found two

horizons with clay-rich diamictic sediments separated by layers of clay and silt. The sediments (except for the upper layer) are normally consolidated. The upper layer is a diamicton that exhibits large variations in texture and in the degree of compactness along the length of the section. Generally, the diamicton is compact where the matrix consists of clay and silt, and normally consolidated where the matrix is sand rich. Four shell dates are obtained; the oldest, yielding 11 970±60 ^{14}C years BP, is from a clay near the base of the section and the youngest, yielding 11 100±60 ^{14}C, is from the upper diamicton (J. Bakke, unpublished data) (Table 1, Figs. 2 and 4). Two of the dates are undertaken on intact, unbroken shells. The section lies at the toe of a very steep fjord side (Fig. 7), and Bakke (pers. comm., 2006) interpreted the upper diamicton as being deposited by gravity flows. According to Bakke no reliable evidence is found in this section to suggest that the site was ice overrun during the YD. On the other hand, lack of overconsolidation does not provide unambiguous proof that the ice did not overrun the site. An incomplete consolidation could be explained by retainment of pore pressure in sediments that were not allowed to drain (see Larsen et al., 1995; Piotrowski & Kraus, 1997).

To conclude, the observations from the outer Hardangerfjorden area are simplest explained in terms of an ice-filled fjord during the YD. Few, if any, of the observations here can be taken to indicate ice-free conditions in the fjord in this period (possibly with the exception of the evidence from the aforementioned section from Tysnes). Even though there is a possibility that the shell-bearing diamictons may have a non-glacial (glaciomarine) origin and/or that some of the moraines may have been formed by land-based ice masses, many observations [including those of Follestad (1972) and Lohne et al. (2003)] remain largely unexplained in the context of an ice-free fjord. This is important to bear in mind when weighing the evidence from the outer Hardangerfjorden area against the findings along the mid and inner parts of the fjord (see also Section 6).

Table 1. Radiocarbon dates undertaken on marine and terrestrial macrofossils from sites along Hardangerfjorden located proximal to the Herdla-Halsnøy moraines (see text). The dates from Nedrevågen, Fjellandsbøvatnet and Skardsvatnet are from Bakke (unpublished data.). The other dates are from the present study. The shell dates from Nedrevågen and Bersenmyra (yielding Bølling/Allerød ages) are corrected for a marine reservoir age of 380 years. The other shell dates (yielding YD ages) are corrected for a marine reservoir age of 610 years (Bondevik et al., 1999). Calibration of the dates was performed with the on-line version of CALIB 5.0 (Stuiver et al., 2005), using both the INTCAL04 (Reimer et al., 2004) and the MARINE04 (Hughen et al., 2004) data sets. Calibrated dates are cited with a 2σ age range.

Locality	^{14}C dates	Laboratory no.	Calibrated age	Material
Nedrevågen, Tysnes (ca. 15-20 m a.s.l.)	11 970±60	Poz-8163	13690-13960	Shell fragments (unidentified)
	11 440±60	Poz-9400	13050-13280	Shell (*Macoma calcarea*)
	11 290±60	Poz-9399	13170-13400	Shell (paired *Mya truncata*)
	11 100±60	Poz-9401	12890-13100	Shell fragments (unidentified)
Fjellandsbøvatnet, Uskedal (191 m a.s.l.)	9980±65	Poz-458	11250-11750	Terrestrial plant macrofossil
Bersenmyra, Gravdal (94 m a.s.l.)	11 515±85	Tua-2838	13210-13570	Shell fragments (unidentified)
Løkjen, Gravdal (78 m a.s.l.)	10 020±80	Beta-142525	11220-11900	Shell fragments (*Hiatella arctica*)
	10 215±80	TUa-2589	11400-12290	Shell fragments (*Mytilus edulis* and *Hiatella arctica*)
Kolltveittjørna, Ljones (82 m a.s.l.)	9995±85	TUa-1551	11200-11880	Shell (*Macoma calcarea*)
	9845±80	TUa-1653	11080-11680	Sea urchin skeletal fragments
Skardsvatnet, Fyksesund (116 m a.s.l.)	9860±50	Poz-590	11200-11390	Terrestrial plant macrofossil

5.4. Discussion of the implications of our hypothesis for the YD ice-sheet limits in the area
The Eidfjord-Osa moraines (Anundsen & Simonsen, 1968) are a part of an extensive moraine system mapped in the inner parts of Hardangerfjorden and Sognefjorden (Liestøl, 1963; Klovning, 1963; Sindre, 1973, 1974; Vorren, 1973; Bergstrøm, 1975; Aa, 1982a, b) (Fig. 2). All the major ice-marginal deposits in the inner branches of Hardangerfjorden, including the Hæreid moraine (see e.g., Monckton, 1899; Kaldhol, 1941; Undås, 1944; Simonsen, 1963; Anundsen, 1964; Kvistad, 1965; Anundsen & Simonsen, 1968; Rye, 1970; Holtedahl, 1975; Hunnes & Anundsen, 1985; Sundal, 1999; Nygaard, 2002), probably belong to this moraine system. Many of the moraines

show evidence of being deposited during periods of glacial advance (e.g., Liestøl, 1963; Anundsen & Simonsen, 1968; Vorren, 1973; Bergstrøm, 1975; Aa, 1982a, b). Several probably also have a polyepisodic origin; initially formed during the deglaciation of the main fjord and modified or rebuilt during a later ice-front readvance. One such example is the moraine in Kinsarvik where stratigraphical investigations indicate that the moraine was modified by a later ice-front readvance (Sundal, 1999). In the adjacent valley, Lofthus (Fig. 1), two sets of moraines occur near the local marine limit, and the youngest show evidence of being formed during an ice-front readvance (Aarseth, 2004, p. 500). On the mountain plateaus, the moraine ridges are often 3-4 in parallel and up to 15 m high (Anundsen & Simonsen, 1968). Judged on the basis of their size, distinctness, distribution and continuity, the moraines represent by far the most marked glacial event(s) in inner Hardanger. In the following, the readvance event is informally referred to as the Eidfjord-Osa event (even though it is probably not strictly synchronous everywhere).

Since the moraines are located inland of the Herdla-Halsnøy moraines they have previously been assigned an early Preboreal age mainly on a morphostratigraphical basis (Anundsen & Simonsen, 1968; Anundsen, 1972; Hamborg, 1983). A ^{14}C date, yielding 9680±90 years BP (11240-10760 cal. years BP), on a branch of juniper (*Juniper communis*) found in a foreset bed in a terrace near the Hæreid moraine was also taken to indicate that the Eidfjord-Osa moraines formed during the early Preboreal (Rye, 1970). However, georadar investigations undertaken in the present study (Paper 2) indicate that the foreset where the juniper branch was found is not an erosional remnant of the original ice-contact delta. Instead, the foreset lies just seawards of a palaeodelta front buried in the moraine, implying that the ^{14}C date only represents a minimum age of the moraine's ice-contact system. It is reasonable to assume that the Eidfjord-Osa event corresponds to the oldest glacial event recorded by Bakke et al. (2005) in the Jondal area, in the mid Hardangerfjorden area (Fig. 1), and which they have attributed to the YD. The Eidfjord-Osa moraines are most likely older than their next glacial readvance event, the `Jondal Event 1`, dated to ca. 11 100 cal. years BP, and which occurred within the `Preboreal Oscillation` (Bakke et al., 2005). Two AMS ^{14}C dates on terrestrial macrofossils from the lake sediments of Skardsvatn and Fjellandsbøvatnet, proximal to the lower-lying moraines along the fjord, yielded ages in the range 11200-11750 cal. years BP, i.e., close to the YD/Holocene transition (Bakke, unpublished data) (Fig. 2; Table 1). These represent minimum ages of the moraines and partly overlap with the dating results of the Herdla-Halsnøy moraines north of Hardangerfjorden (ca. 11 600-11 700 cal. years BP; Bondevik & Mangerud, 2002).

If Hardangerfjorden remained ice free during the YD, this probably implies that the Eidfjord-Osa moraines include moraines that were formed during the YD ice-front readvance, or that were formed during the Bølling/Allerød deglaciation and later modified or rebuilt during this readvance. Since there is no reason to call in question the published reconstructions of the YD ice margin in the areas to the north and to the south of the fjord inlet, this implies (if our hypothesis is correct) that Hardangerfjorden constituted an open ice-free corridor at the time the Herdla moraine, the Ulven moraine at Os, and the moraines in Etne (Fig. 2) were formed. As already noted, a scenario with Hardangerfjorden constituting an open ice-free corridor surrounded by land areas that were largely covered by ice during the YD resembles the ideas put forward by the early geoscientists working in the area. Kolderup (1908) correlated the major ice-marginal deposits in Eidsfjorden, at the southwestern inlet of the Stølsheimen mountain plateau (Fig. 2), to the Ra (YD) Moraines in eastern Norway. Kaldhol (1941) and Undås (1944) correlated the major ice-marginal deposits at the head of Hardangerfjorden, including the Hæreid moraine, to the same moraines. Simonsen (1963) supported their view, although he did not rule out the possibility that the moraine system is of early Preboreal age. An important argument in this respect was that the lowering of the firn limit estimated for the Eidfjord-Osa event was comparable to that inferred for the YD ice readvance (Simonsen, 1963).

From a glaciological perspective, the idea of an ice-free corridor is not unreasonable, given that the winter precipitation in the area during the YD probably was considerably less than the present (cf. Dahl & Nesje, 1992; Bakke et al., 2005), implying that the ice production on the mountain areas surrounding the fjord may have been too low to allow a build up of a major tide-water glacier in the fjord. In addition, a strong precipitation gradient (stronger than the present, where the precipitation a few tens of km inland of the extreme west coast is many times that of the interior; Fig. 1) is likely to have existed across western Norway during the YD. Ice caps and ice domes that were situated on the landblock north of Hardangerfjorden (cf. Skår, 1975; Aa & Mangerud, 1981; Hamborg & Mangerud, 1981), and on the Folgefonna peninsula south of the fjord (cf. e.g., Follestad, 1972; Bakke et al., 2005), prior to and during the YD, may effectively have starved the eastern areas of precipitation, and thus retarding glacier growth in inner Hardanger. However, as discussed below, there is presently no observational basis to suggest a direct connection between the Eidfjord-Osa moraines and the YD moraines respectively north (the Herdla moraine, the Ulven moraine and others) and south (including the moraines in Etne) of the fjord.

In inner Hardanger, the system of more or less continuous moraines ends rather abruptly west of Ulvik (Fig. 2). The lack of similar, well-defined ice-marginal deposits in the Voss-Granvin area and further westwards is remarkable. Originally, the moraines in Ulvik were correlated with the moraines in Flåm, in inner Sogn, via the ice-marginal deposits in Raundalen, east of Voss (Anundsen & Simonsen, 1968; Anundsen, 1972) (Fig. 2). A correlation via the deposits in Raundalen was later disputed (e.g., Vorren, 1973; Hamborg, 1983). Hamborg (1983) suggested instead that the deposits at Bolstadøyri, west of Voss (cf. Skreden, 1967), and the ice-marginal deposits in Eidsfjorden, at the southwestern inlet of the Stølsheimen mountain plateau (cf. Aa & Mangerud, 1981) (Fig. 2), correspond to the Eidfjord-Osa moraines. Hamborg (1983) indicated thereby that the ice cap(s) on the landblock between Sognefjorden and Hardangerfjorden were dynamically in connection with the inland ice at the time of formation of the major ice-marginal deposits in the inner branches of Hardangerfjorden. This interpretation was also adopted by Anundsen (1985) who indicated a similar ice cover (albeit somewhat less extensive than suggested by Hamborg) on the landblock between Sognefjorden and Hardangerfjorden during the Eidfjord-Osa event (cf. Fig. 14 in Anundsen, 1985).

Sindre (1974) suggested a correlation between the Eidfjord-Osa moraines and the Holmo-Bakka moraines in Nærøyfjorden, at the head of Sognefjorden (Fig. 2). Observations of glacial striations indicate that the Holmo-Bakka moraines were deposited by an outlet glacier from an ice-accumulation centre near the Vikafjell-Fresvikbreen area (Sindre, 1974) (Fig. 2). A number of moraines attributed to the Eidfjord-Osa event have been mapped near continuously in the northern, central Stølsheimen area (Aa, 1982b) (Fig. 2). The same centre that nourished the outlet glacier responsible for the Holmo-Bakka moraines nourished probably also the outlet glaciers responsible for the easternmost of these moraines (cf. Aa, 1982b). These observations add support to the notion that large ice cap(s) existed on the landblock between Sognefjorden and Hardangerfjorden which were dynamically in connection with the inland ice during the Eidfjord-Osa event.

Along the northwestern side of Hardangerfjorden, traces of former ice-margins occur mainly as isolated moraine ridges and other proglacial deposits in sidefjords and tributary valleys (Fig. 2). A terminal moraine occurs near the mouth of Fyksesundsfjorden (Rekstad, 1911; Holtedahl, 1975; Hamborg, 1983; Hoel, 1992) and in Øystese (Holtedahl, 1975) (Fig. 2, see Fig. 1, for location). Some of the highest terraces upvalley of Norheimsund (Fig. 1; cf. Neteland in Fig. 3) are probably remnants of a proglacial fan that was deposited by meltwater from a glacier in the Kvamskogen valley (Hamborg, 1983). A proglacial, terraced deposit also occurs near Strandebarm (Fig. 1), defining the marine limit there (Hamborg, 1983) (Fig. 3). Observations of glacial striations and other ice-directional elements indicate that individual ice-accumulation centres have existed in the Botnavatn area, and probably also in the mountain areas between Kvamskogen and Bergsdalen during a late stage of the last glaciation (Hamborg & Mangerud, 1981) (see also Section 5.5).

Outlet glaciers descending from these ice-accumulation centres were probably responsible for the moraines and proglacial deposits in the lowland areas (Fig. 2).

   There is, at present, no firm evidence to support or negate a correlation of these moraines and proglacial deposits. Hamborg (1983) attributed the moraines in Fyksesundsfjorden and in Øystese, and the proglacial deposits at Neteland and in Strandebarm to different shorelines (bold letter types in Fig. 3). He indicated, on the basis of his shoreline reconstructions, that the moraine in Øystese is the youngest of these, about 200 (^{14}C) years younger than the terraced, proglacial deposit in Strandebarm. However, it should be stressed that Hamborg`s (1983) shorelines are assigned an absolute age mainly on the basis of regional interpolations. Marine shells (*Mytilus edulis*) found in a terrace with the surface ca. 68 m a.s.l. in Øystese are dated to 9420±130 ^{14}C years BP (9920-10550 cal. years BP) (B.G. Andersen pers. comm. in Hamborg, 1983). This date does not fit well into Hamborg`s shoreline reconstruction as his `9400-shoreline` lies some 25 m below the site where the dated shells were found (Fig. 3). Moreover, the exact sea level during the formation of the moraines in Fyksesundsfjorden and Øystese is uncertain. In addition, there is not always agreement among the authors on what should be considered as the marine limit in an area. Holtedahl (1975) indicated a marine limit at the head of Fyksesundsfjorden of about 80 m a.s.l., whereas Rekstad (1911) and Hamborg (1983) estimated it to be about 90-94 m a.s.l. Our estimates of the marine limit at Ljones (ca. 98-99 m a.s.l.; Paper 3) and at Gravdal (ca. 94 m a.s.l.; Fig. 11; Paper 4) are considerably higher than Hamborg`s (1983) estimate for the same areas (Fig. 3). If, as proposed in the present thesis, several of the highest shore levels along the middle and inner parts of Hardangerfjorden can be attributed to the same (YD maximum highstand) shoreline, the terminal moraines in Fyksesundsfjorden and in Øystese, and the proglacial deposits at Neteland and in Strandebarm could in principle belong to the same glacial event. A minimum age of the deglaciation of Fyksesundsfjorden (and of the moraine here) is provided by an AMS date, yielding 9860±50 ^{14}C years BP (11200-11390 cal. years BP) on a terrestrial macrofossil from the lake Skardsvatn, located on a ledge along the western side of the fjord (Bakke, unpublished data) (Table 1; Fig. 2). This suggests that the moraine in Fyksesundsfjorden has a closely similar age as several of the Eidfjord-Osa moraines and as several of the Herdla-Halsnøy moraines (Fig. 4).

As noted above, the moraines in the Os-Fusafjorden area (including the Ulven moraine) could have formed independently of whether there was ice in Hardangerfjorden or not during the YD. Ice accumulation centres located in the Gullfjellet area, east of Bergen, and in the Botnavatn area, east of Samnangerfjorden (Fig. 1) during the YD (Skår, 1975; Hamborg & Mangerud, 1981), nourished the branch of the ice sheet responsible for these moraines. They could thus in principle (if our hypothesis is correct) be contemporaneous with the Eidfjord-Osa moraines. The moraines on Tysnes, and those between this island and the island Halsnøy, cannot however be contemporaneous with the Eidfjord-Osa moraines and must have been formed earlier, by a glacier in Hardangerfjorden. Nor is it likely that the moraine in Bjørnafjorden and the moraines on Halsnøy are contemporaneous with the Eidfjord-Osa moraines, although there is a possibility that some of the deposits directly proximal to these moraines [e.g., the sedimentary ridge at Ølve (Fig. 10) and the moraine ridges of the Lunde stage (Aarseth, 1971)] may have formed after the main fjord became deglaciated. This would in case require a significant local ice production in the outer coastal areas, including in the lowland areas, probably implying an east-west precipitation gradient that was considerably stronger than the present (see above).

   In conclusion, it is not known how far west the ice front extended on the landblock north of Hardangerfjorden during the Eidfjord-Osa event. A situation with large glaciers still remaining in the tributary valleys on the northern side of the fjord during this event [implying that the landblock was more extensively covered by ice during this event than that indicated by Hamborg (1983)] is possible in the context of previous findings. However, a correlation between the Eidfjord-Osa moraines and the moraines along the northern side is highly uncertain. Also a correlation between moraines along the fjord and the Ulven moraine remains unsubstantiated (even though the moraines
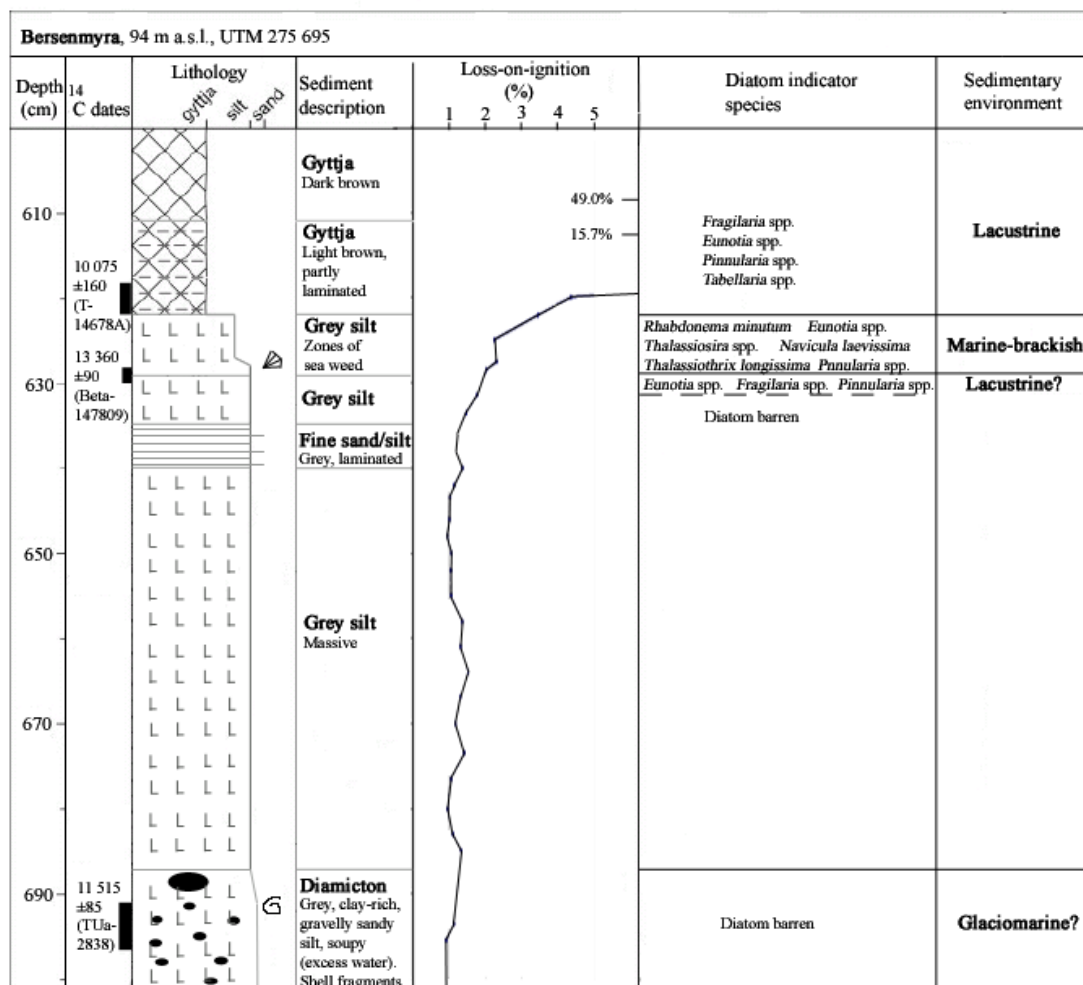
Bersenmyra, 94 m a.s.l., UTM 275 695						
Depth (cm)	14 C dates	Lithology (gyttja, silt, sand)	Sediment description	Loss-on-ignition (%) 1 2 3 4 5	Diatom indicator species	Sedimentary environment
610			**Gyttja** Dark brown	49.0% —		**Lacustrine**
	10 075 ±160 (T-14678A)		**Gyttja** Light brown, partly laminated	15.7% —	*Fragilaria* spp. *Eunotia* spp. *Pinnularia* spp. *Tabellaria* spp.	
	13 360 ±90 (Beta-147809)	L L L L	**Grey silt** Zones of sea weed		*Rhabdonema minutum* *Eunotia* spp. *Thalassiosira* spp. *Navicula laevissima* *Thalassiothrix longissima* *Pnnularia* spp.	**Marine-brackish**
630		L L L L L L L L	**Grey silt**		*Eunotia* spp. *Fragilaria* spp. *Pinnularia* spp. Diatom barren	**Lacustrine?**
			Fine sand/silt Grey, laminated			
650		L L L L L L L L L L L L L L L L L L L L				
		L L L L L L L L	**Grey silt** Massive			
670		L L L L L L L L L L L L L L L L				
		L L L L L L L L				
690	11 515 ±85 (TUa-2838)		**Diamicton** Grey, clay-rich, gravelly sandy silt, soupy (excess water). Shell fragments		Diatom barren	**Glaciomarine?**

Fig. 11. Lithostratigraphy, loss-on-ignition, radiocarbon dates and diatoms in Bersenmyra, Gravdal. The interpretation of the upper silt layer as marine-brackish is based on the findings of sea weed fragments, and on the occurrence of marine and brackish water diatoms, together with freshwater diatoms, in this interval. During the maximum highstand, sea level was probably close to the basin threshold. The two upper dates (T-14678A and Beta-147809) are undertaken on bulk gyttja samples, whereas the lower date (TUa-2838) is undertaken on shell fragments (Table 1). The oldest bulk date is clearly erroneous. Since the bulk samples probably contain aquatic organic matter, the dates may include an (unknown) reservoir age (cf. e.g., Olsson, 1991). The shell date yielding Allerød age at the base is corrected for a reservoir age of 400 years (Tabell 1). Note however that neither the Vedde Ash Bed nor organic-enriched Allerød sediments are found in the deposits, implying that the shell date may be too old. Owing to its loosely compacted (soupy) nature, the basal diamicton is interpreted to be of glaciomarine origin. However, the possibility that the diamicton is a till cannot be completely ruled out.

could in principle be contemporaneous). Such a correlation would in case leave many observations in the outer part of the fjord, including those made by Lohne et al. (2003), unexplained (see above).

On the opposite (southern) side of the fjord, there is evidence in support of significant land-based glaciation during the Eidfjord-Osa event, probably with glaciers remaining in the tributary valleys along the fjord, including in the outer part. In inner Hardanger, the Eidfjord-Osa moraines are correlated with moraines at the head of Åkrafjorden (Fig. 2), and, although a direct connection is yet to be established, with the Trollgaren moraines further to the south (Anundsen, 1972; Hamborg, 1983; Anundsen, 1985). According to this reconstruction, the ice cap on the Folgefonna mountain plateau was not dynamically connected to the inland ice during the Eidfjord-Osa event. Nevertheless, it is likely that traces of this event are present also on the Folgefonna peninsula, outside the present-day ice cap. The distinct (several hundred-meter-long) moraine ridge south of Aga (Kvistad, 1965) (Fig. 2) may well represent the Eidfjord-Osa event. The moraine was deposited by valley glacier(s) descending from the northern part of the Folgefonna mountain plateau and is correlated with the moraine in Odda (Kvistad, 1965).

Also on the opposite side of the Folgefonna peninsula there are probably traces of this event (Fig. 12). By combining the position of marginal moraines with ^{14}C dated meltwater deposits in proglacial lakes, a continuous high-resolution record of variations in glacier size and equilibrium-line altitudes in the Jondal area during the Lateglacial and early Holocene has been obtained (Bakke et al., 2005). The oldest glacial event recorded was attributed to the YD (Bakke et al., 2005) and corresponds most likely to the Eidfjord-Osa event. The marginal moraines at Stormyr and Drebrekke (Fig. 12), which formed by low-altitude cirque glaciers during this event, were attributed to the YD based on the following arguments (Bakke et al., 2005): (1) the radiocarbon-based chronology suggests a YD age of the oldest meltwater-induced sediments, (2) the lowering of the equilibrium-line altitude is substantial, estimated to ca. 355 m below the present, and (3) the estimated mean ablation-season temperatures and winter precipitation values point to a YD age of formation. If Bakke et al.`s (2005) reconstructions are correct, this supports our hypothesis of an ice free Hardangerfjorden during the YD. Mangerud (2000), on the other hand, suggested that the low-lying moraines at Stormyr and Drebrekke instead were formed during the early Preboreal,
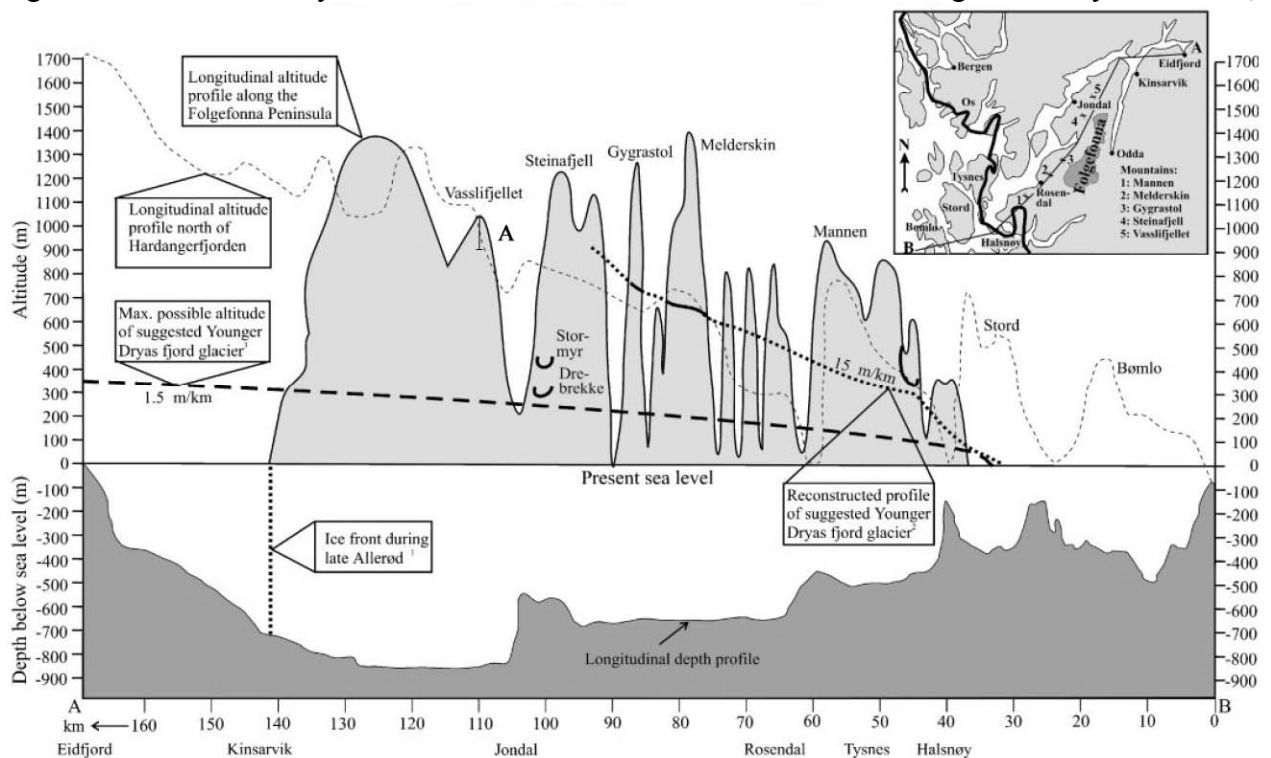
Fig. 12. Suggested YD fjord glacier scenarios along Hardangerfjorden, according to Bakke et al. (2005). The upper bold dotted line shows the reconstructed YD fjord glacier after Follestad (1972), whereas the lower bold punctuated line marks the highest possible ice sheet/fjord glacier based on the YD cirque glacier at Drebrekke in Jondal. The height interval where *Artemisia norvegica* (A) occurs on Vasslifjellet is also indicated.

by ice masses left on the land areas after the main fjord was deglaciated.

In Bondhusdalen, south of Maurangerfjorden, an end moraine lies on top of a marine terrace attaining an altitude of ca. 100 m a.s.l. (the marine limit) (Rekstad, 1907; Hunnes & Anundsen, 1985; Riis Simonsen, 1999). The moraine was probably deposited by an outlet glacier that readvanced and overrun the southern (upvalley) part of the terrace (Hunnes & Anundsen, 1985). Also in the neighbouring valleys around Maurangerfjorden, there are moraines and proglacial deposits located close to the marine limit, demonstrating that many of the tributary valleys were occupied by dynamically active glaciers after the main fjord was deglaciated. There is, at present, no firm evidence to support or negate a correlation of these moraines and proglacial deposits with the Eidfjord-Osa moraines. Based on Hamborg`s (1983) shoreline reconstructions, the terrace in Bondhusdalen was formed about 350-400 (^{14}C) years before the moraine in Eidfjord. However, if,

as proposed in the present thesis, several of the highest shore levels along the middle and inner parts of Hardangerfjorden developed during the maximum highstand of the YD transgression, the low-lying end moraines in Bondhusdalen, and in several of the neighbouring valleys around Maurangerfjorden, could in principle be age equivalent with the Eidfjord-Osa moraines.

As earlier mentioned, on the southern part of the Folgefonna peninsula, there is evidence suggesting that large glaciers still remained in the tributary valleys after the deglaciation of the main fjord (Follestad, 1972). On Snilstveitøy and in the bay inside this island (Fig. 7), observations of glacial striations and proglacial deposits indicate that a piedmont-like glacier established with a more or less stationary calving front, and which later broke up into separate valley glaciers (Follestad, 1972). Moreover, a distinct moraine ridge system occurs directly east of the Herdla-Halsnøy moraines (Fig. 2), representing a southwestern branch of the Folgefonna glacier (Follestad, 1972). The moraine system is attributed to an ice-front readvance event termed the Blådalen substage and tentatively correlated with the Eidfjord-Osa event (Follestad, 1972).

Altogether, the observations allow for the possibility that Hardangerfjorden constituted an ice free corridor surrounded by land areas largely covered by ice during the Eidfjord-Osa event. However, there is presently no basis for correlating the Eidfjord-Osa moraines (or any of the other moraines along the fjord, inside Huglo-Halsnøy) with the YD moraines north and south of the fjord inlet. Such a correlation would not only be in conflict with the observations indicating (if correctly interpreted) ice-overrun Allerød and YD marine sediments in the outer part of Hardangerfjorden, but would also be in conflict with previous reconstructions of the ice margin during the Eidfjord-Osa event south of Hardangerfjorden (where the Eidfjord-Osa moraines are correlated with the Trollgaren moraines, see Anundsen, 1972). However, to this it should be added that there is a considerable degree of uncertainty attributed to the determination of which moraines south of Hardangerfjorden that correspond *in time* to the Herdla-Halsnøy moraines. This is because (1) the YD ice readvance(s) between Hardangerfjorden and Boknafjorden generally are dated to the early-to-mid-YD period, rather than to the end of this period (Anundsen, 1972; Anundsen, 1977b; Blystad & Anundsen, 1983; Andersen et al., 1995), and (2) the two moraine systems proximally to the YD moraines here, belonging respectively to the Trollgaren and the Blåfjell substage, are not age constrained by direct dating evidence but have been assigned a Preboreal age mainly on a morphostratigraphic basis (e.g., Anundsen, 1985). [The Herdla-Halsnøy moraines probably correspond in time to the Ås-Ski moraines in the Oslofjorden area, in eastern Norway, and not to the somewhat older Ra moraines (Mangerud, 2004). If correct, this means that the moraines corresponding to the Ås-Ski moraines had to cross the moraines corresponding to the Ra moraines somewhere between Oslofjorden and Hardangerfjorden to connect with the Herdla-Halsnøy moraines. This crossing point is yet to be identified (Mangerud, 2004)].

5.5. Does the present-day occurrence of Artemisia norvegica in the mountains imply a thin ice during the YD?

On Vasslifjellet in Jondal (Figs. 1 and 13), an abundant population of *Artemisia norvegica* ('norsk malurt') has been documented (Moe et al., 1994; Moe, 2004). This is a plant of great phytogeographical significance, occurring today in geographically disjunct and well-circumscribed areas (Fig. 13). They are commonly interpreted as relics of earlier populations that over wintered (at least through the YD) on nunataks (see Paus et al., in press). The population of *A. norvegica* on Vasslifjellet is located within a small, well-defined area between 900 - 1030 m a.s.l. Moe et al. (1994) indicated that it may be a relic of a larger population from Pleniglacial time which has survived the YD on a local nunatak. He argued that its disjunct geographical distribution (*A. norvegica* is not found e.g., on Hardangervidda) makes it less likely that the species followed the retreating ice front after the YD glaciation. If true, i.e., that the surface of the YD ice sheet did not exceeded 900 m a.s.l. at this locality, this has important implications, not only for the ice height reconstructions, but also for the reconstructions of the YD ice movements in the area.

According to the reconstruction of Hamborg & Mangerud (1981), the sites where *A. norvegica* has been found were covered by 300-500 m thick ice during the YD glaciation maximum. Hamborg & Mangerud (1981) based their ice height reconstruction on observations of lateral moraines correlated with the Herdla-Halsnøy moraines, on observations of glacial striations and on theoretical calculations of ice profiles. The lateral moraines on the mountains Gygrastolen and Melderskin, and some steeply dipping moraines in the outer Hardangerfjorden area (Follestad, 1972;Fig. 12), formed the basis for their ice-surface reconstruction along the south side of the fjord.

Fig. 13. The present-day occurrence of *Artemisia norvegica* in Eurasia. From Helland-Hansen (2004).

A reduction of the glacier surface in the middle Hardangerfjorden area by 300-500 meters or more implies that the lateral moraines along Hardangerfjorden (at least that on Gygrastolen) are probably not from the YD, but date back to an earlier period. A reduction of the glacier surface would also imply an ice flow during the YD that was more topographically dependent than that envisaged by Hamborg & Mangerud (1981) for the area. On the northern side of Hardangerfjorden, Hamborg & Mangerud (1981) attributed their 'younger regional ice movements' to the YD period, representing an ice flow directed from the fjord towards Kvamskogen and Samnangerfjorden in the west (Fig. 14). Their reconstruction implies that the YD ice drained from an area of the fjord which at the present has a water depth of more than 800 m [plus an additionally 200-250 m of fjord sediments (Hoel, 1992)] towards west across mountains attaining altitudes of more than 900 m a.s.l. Clearly, this would not be possible if the ice height was less than 900 m a.s.l. during the YD. This could mean that several of the glacial striations belonging to Hamborg & Mangerud`s (1981) 'younger regional ice movements' date from an older, pre-YD period, and that striations from the YD ice are included in their `youngest ice movements` (Fig. 14). The latter movements reflect outlet glaciers descending from ice accumulation areas centred in the Botnavatn-Kvamskogen-Bergsdalen area and which were responsible for several of the proglacial/ice marginal deposits in the middle Hardangerfjorden area [including the proglacial deposit in Strandebarm (Hamborg, 1983), and the end moraines in Øystese and in Fyksesundsfjorden (e.g., Holtedahl, 1975) (see Section 5.4)]. A re-interpretation of the ice flow pattern here would also have implications for the correlation and interpretation of glacial striations in the other, neighbouring mountain areas [cf. e.g., Mæland`s (1963) and Skreden`s (1967) southerly to south-easterly directed striations in the mountain areas between Granvin and Voss (Fig. 1)]. Moreover, if the *A. norvegica* populations are relics of earlier populations that over wintered through the YD on nunataks, this would also have implications for the reconstructions of longitudinal glacier profiles in the neighbouring fjords (see e.g., Andersen et al, 1995). However, the possibility that the site where *A. norvegica* is presently growing was overrun by the YD ice, and that these plants immigrated during the early Holocene,

cannot be ruled out. If so, then Vasslifjellet in Jondal would be the only known *A. norvegica* locality in North Europe that has been overrun by an ice sheet during the YD.
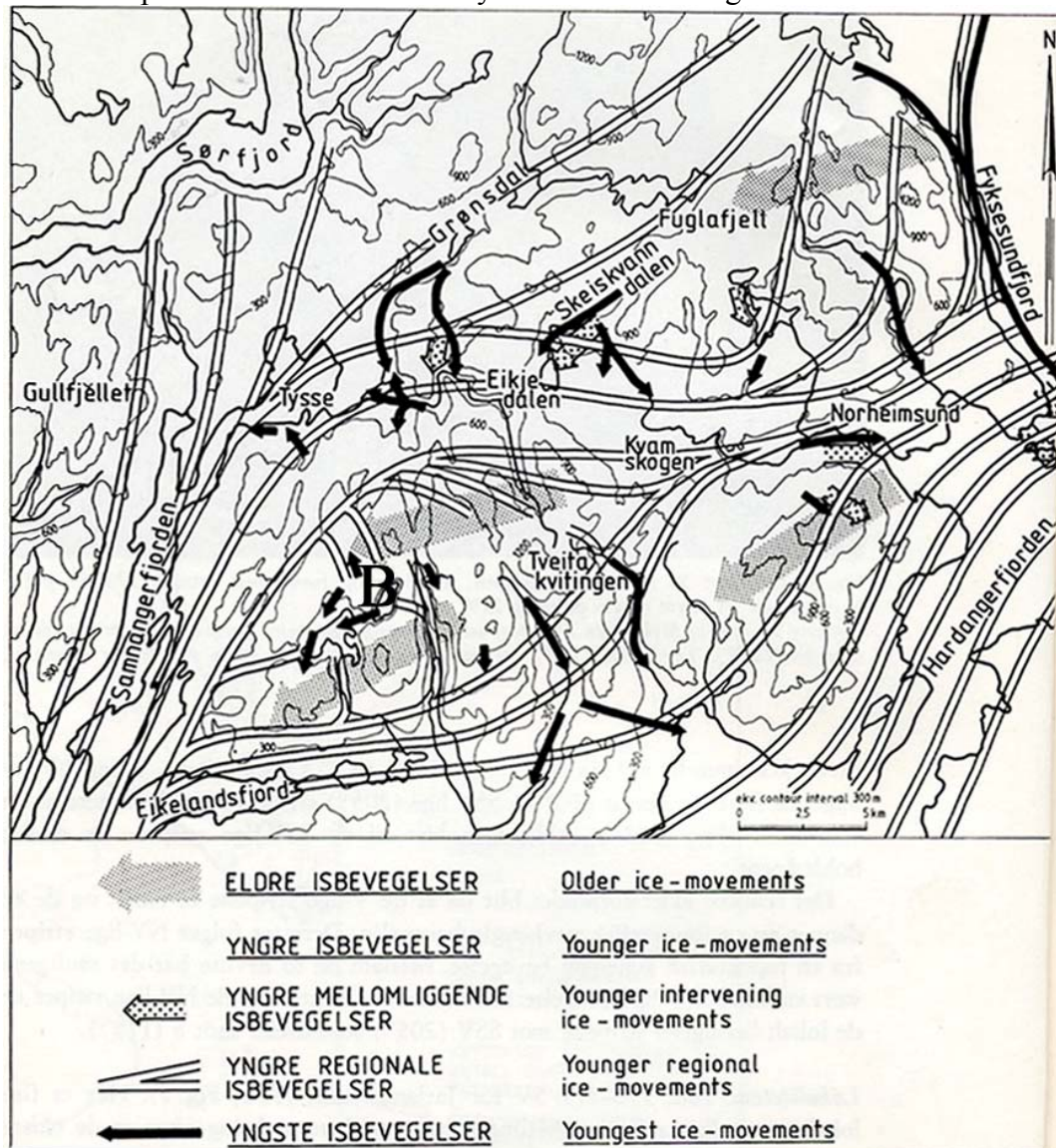

Fig. 14. Ice movement directions on the northern side of Hardangerfjorden, according to Hamborg & Mangerud (1981). For location, see Fig. 1. B – the Botnavatn area.

6. A critical evaluation of our interpretations of the sea-level records along Hardangerfjorden

As discussed above, the observations from the outer fjord area are very difficult to reconcile with the hypothesis of an ice free fjord during the YD. The findings along the mid and inner parts of the fjord suggesting ice free conditions must therefore be weighted against these observations. Since the absolute age control on the timing of the deglaciation and the earliest sea levels is poor in the present work (due to diverging radiocarbon dates and lack of unequivocal time synchronic markers in the lower stratigraphic columns), the hypothesis of an ice free Hardangerfjorden during the YD mainly relies on a correct interpretation of the sea-level records along the fjord (i.e., that they are correctly interpreted as the product of a RSL cycle).

In seeking a solution to this conflict, I here discuss alternative interpretations of our sea-level data. Alternative interpretations have previously been proposed by Mangerud (2000), and his re-interpretations will be included in the discussion below. Several of the alternatives listed below have also (to some extent) been discussed in the papers in the present thesis:

-The RSL cycle recorded postdates the YD ice advance.
-The sediments in the isolation basins are reworked and/or there is a major hiatus in the stratigraphic records.
-The biostratigraphic records in the isolation basins are incorrectly interpreted; RSL fall was monotonic (cf. Mangerud, 2000).
-The stratigraphy of the Hæreid moraine is incorrectly interpreted and cannot be used to estimate sea level.

6.1. The RSL cycle recorded postdates the YD ice advance.

At the outer coast, two RSL cycles of similar magnitude as that recorded in the present study are known; the YD transgression, which took place in the late Allerød-YD interval (e.g., Lohne et al., 2004), and the mid-Holocene (Tapes) transgression (e.g., Kaland, 1984). The RSL cycle we have postulated along Hardangerfjorden is definitely older than the Tapes transgression. It is also highly unlikely that the RSL cycle dates from the period between these two RSL events. The early Preboreal is a period characterised by extremely high uplift rates, particularly in the fjord-head areas. This would in case imply that the inner-to-mid parts of the fjord experienced a RSL rise of 10 to 15 m at the time the RSL was dropping fast in the outer coast! Therefore, the possibility that the RSL rise dates from the end of the YD or from the Holocene can almost certainly be ruled out.

6.2. The sediments in the isolation basins are reworked and/or there is a major hiatus in the stratigraphic records.

Reworking is an important factor that requires attention when interpreting the isolation basin data. Clearly, if the interpretation is based on observations on reworked deposits, this could lead to an incorrect reconstruction of the sea level history. In the present study, the main depositional units in each basin were mapped in the field through detailed profiling with a Russian peat corer and a 110 mm piston corer. Parallel cores were examined for diatoms to verify the observations from the main core. In this way, we were able to ascertain that important horizons, such as the marine layer in Vambheim-119 (Fig. 15a), the lower freshwater layer in Bu-113 (Fig. 15b), the brackish water horizon in Tveitatjørna (Fig. 6 in Paper 3), and the organic-enriched layer J in Mjådalsmyra (Fig. 11 in Paper 3) are present as discrete horizons in the central parts of the respective basins. [Note however that in Mjådalsmyra we were only able to penetrate the deepest parts of the stratigraphy with a few 110mm cores whereas the Russian corer stopped higher up in the stratigraphy (Fig. 11 in Paper 3)]. Deformation features or other evidence of slumping were not revealed at critical stratigraphic levels in the cores. The succession of biostratigraphic units in the basins Vambheim-119 (where two freshwater units are separated by a marine unit; Fig. 15a), Tveitatjørna (where two freshwater units are separated by a brackish water unit) and Mjådalsmyra (where two marine units are separated by a brackish water unit) cannot be explained simply as a result of reworking from the underlying basin sediments. This is because (1) the lithology in the lower unit (organic-poor, minerogenic sediments) is markedly different from that in the upper unit (gyttja), (2) the number of microfossils is significantly higher in the upper unit, and (3) the degree of preservation of the microfossils in the upper unit is at least as good as in the lower unit. Also in Bu-113 (Fig. 15b), where the two marine units have a similar lithology (mainly sandy silt), it is unlikely that the occurrence of marine diatoms in the upper unit is caused by reworking of frustules from the underlying units. Despite low diatom counts, the number of intact frustules appears to be higher in the upper unit than in the lower unit and the spectra in the upper unit do not exhibit a heterogenous mixture of freshwater and marine species, as one might expect if the sediments were reworked. (The interpretation of the diatom stratigraphy is discussed more thoroughly in the next section.). Nor is it likely that the upper marine unit has been emplaced over younger units e.g., by slumping, given the lateral extent of the unit and the lack of evidence of sediment disturbances.
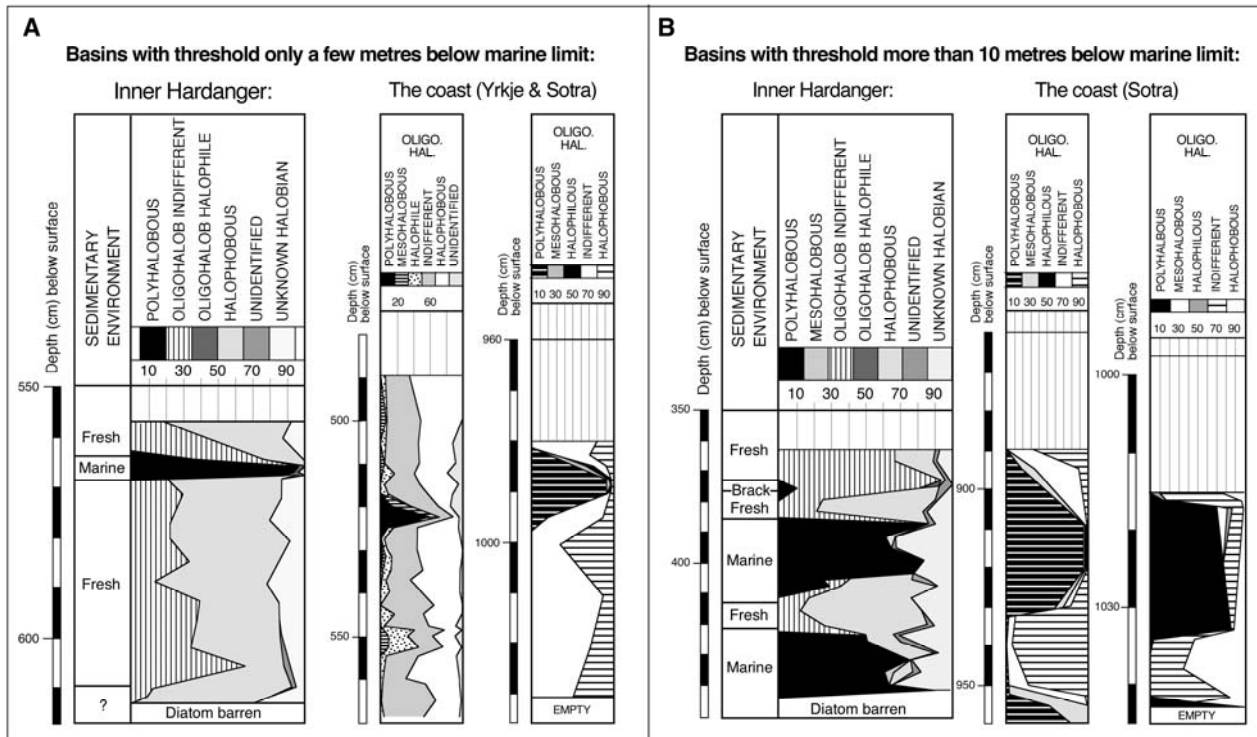
Fig. 15. A comparison with diatom records from localities at the outer coast documenting the YD transgression. A. Total diagram for diatoms from (left to right): Vambheim-119 (Helle et al., 1997), Meåstjørna (Braaten & Hermansen, 1985) and Kvernavatn (Krzywinski & Stabell, 1984). The last two basins are from Yrkje and Sotra, respectively (see Fig. 1, for location). B. Total diagram for diatoms from (left to right): Bu-113 (Helle et al., 1997), Klæsvatn and Herøyvatn (Krzywinski & Stabell, 1984). The last two basins are from Sotra. The basins are situated a few metres below (A) and more than 10 m below (B) the local marine limit, respectively. Note that there are low diatom counts in the depth interval below 385 cm in Bu-113. From Helle et al. (1997).

However, this does not rule out the possibility that some reworked microfossils and other remains are embedded in the deposits. In none of the cores investigated, the Vedde Ash is present as a well defined, visible bed. The low concentrations of Vedde Ash and the fact that rhyolitic glass shards with the same major element chemistry as the Vedde Ash are found also well below the level of maximum shard concentration in some of the basins, including in Bu-113 (Paper 1) and Mjådalsmyra (Paper 3), may indicate that the entire tephra sequence consists of redeposited ash grains and thus postdates the Vedde Ash fall (cf. Mangerud, 2000). In Mjådalsmyra, the level of maximum peak occurrence falls within the interval of an organic enriched layer (layer J; Fig. 12 in Paper 3). This gives ground for suspicion that this sediment layer may contain a mixture of redeposited material, even though the peak occurs where the loss-on-ignition values are declining (see discussion in Paper 3). Redeposition is a potential source of error that could make the macrofossil (seaweed) ^{14}C date from this interval considerably older than the enclosing sediments (Fig. 12 in Paper 3). On the other hand, the anomalous young bulk date in the same interval makes redeposition seem less likely. If the sediments contain substantial amounts of reworked material, one might expect that the bulk dates were biased towards older ages by the redeposited, older organic matter. In addition, the well-preserved pollen grains and the lack of pre-Quaternary microfossils further indicate little redeposition for this part of the stratigraphy (although pre-Weichselian microfossils cannot be excluded).

The occurrence of rhyolitic shards below the maximum peak occurrence in Bu-113 and Mjådalsmyra could also be the result of bioturbation. In some studies, the presence of Vedde Ash below the maximum concentration has been attributed to downward relocation of shards by plant roots penetrating the layer containing the ash (Davies et al., 2005). Rhyolitic shards with the same major element chemistry as those in Vedde Ash may also have been erupted one or several times before the eruption of the Vedde Ash itself (Bond et al., 2001). If this is the case, the occurrence of

Vedde-like rhyolittic ash in the lower part of the stratigraphy may also include redeposited material from a volcanic event that predates the eruption of the Vedde Ash. The horizon with maximum shard concentration in our cores may thus correlate to the Vedde Ash Bed and the tail above represent Vedde Ash that is redeposited, either from shallow areas of the basin or from areas outside the basin. In Tveitatjørna, the sharp definition of the ash peak is inconsistent with reworking, although the shard counts are too low to conclude with confidence that the ash is of primary origin (see discussion in Paper 3).

Another problem concerns the possibility that the stratigraphic records contain hiatuses. The absolute dating control is too weak to rule out the possibility that minor hiatuses are present in the stratigraphic record. In many basins, the changes in the aquatic flora are abrupt and are often accompanied by sharp lithological boundaries. During the isolation (ingression) process, a hiatus may develop between the brackish and marine sediments as tidal currents across the basin threshold erode sediments. The sharp lithological boundaries seen in some of the investigated basins (e.g., in Bu-113) may thus constitute minor erosional hiatuses. Often, however, the abrupt change in the aquatic flora suggests a rapid shift in depositional environment (e.g., from marine to lacustrine), and do not necessarily imply erosion of the basin deposits. In basins that have a sheltered aspect towards the fjord (e.g., Mjådalsmyra), the change may be more abrupt than in basins with an open aspect (e.g., the 89-m basin), because input of sea spray in the latter case has maintained slightly brackish conditions also after the basin had emerged from the sea, resulting in a more gradual change in the aquatic flora (see Paper 3). A permanent lake-ice cover may result in a non-depositional hiatus or in strongly reduced sedimentation rates in lacustrine intervals (cf. the high-altitude lake studied by Paus et al., in press) (Fig. 17; Section 6.3, below).

It is however unlikely that the basin deposits contain major (>500-1000 years) hiatuses resulting from erosion by a glacier (except perhaps for the base of the cores where diamict or deformed sediments are revealed). For example, the loss-on-ignition curves from Bu-113 and Vambheim-119 can be correlated in great detail, even in the lower parts of the stratigraphy (Helle, 1993). Also the pollen stratigraphy in the investigated basins, including the three upper basins at Ljones, shows a similar development. To conclude, the stratigraphic records are not considered to be significantly disturbed by redeposition or to contain major hiatuses.

6.3. The biostratigraphic records in the isolation basins are incorrectly interpreted; RSL fall was monotonic (cf. Mangerud, 2000).

The biostratigraphy leaves little doubt that the upper isolation basins investigated in the present study experienced major shifts in the basin water salinity during the early post-deglaciation time, prior to the onset of the early Holocene sea level regression (Fig. 15). In general, we have attributed these shifts to a major RSL cycle that occurred superimposed upon the overall RSL fall immediately following the deglaciation. However, the interpretation of the palaeoenvironment is complicated by the fact that the investigated sites are located in a fjord environment, several tens of km inland of the outer coast. Here, strong (seasonal and perennial) changes in the sea surface salinity are likely to have occurred in the past, in particular during the deglaciation, due to fluctuations in the upland freshwater discharge. The question is thus whether the shifts observed are unrelated to an ingression/isolation process, and that the basin water salinity instead was determined by the amount of fresh water inflow into the fjord, implying that, even in basins fully connected to the sea, there was a dominance of fresh water aquatics and phytoplanktons in the water column. The latter interpretation was proposed by Mangerud (2000) to explain the stratigraphy in the basins Vambheim-119 and Bu-113 (Paper 1). He suggested that the freshwater diatoms in the lower, minerogenic part of the basin stratigraphy of Vambheim-119 and Bu-113 (i.e., in the depth intervals 569-612 cm and 407-420 cm, respectively; Fig. 15a, b) were brought into the fjord with (meltwater) rivers and deposited in a marine or glacimarine environment.

However, I find this freshwater-discharge hypothesis difficult to reconcile with the following observations:

i) Marine diatoms are totally absent in the majority of diatom spectra in the lower freshwater zones (i.e., in the sediments Mangerud has re-interpreted as marine- or glacimarine). In contrast, the marine diatom zones are dominated by salt-demanding (polyhalobous) species, indicating a marine littoral environment. (Note however that in Bu-113, the number of diatoms counted is low in the interval below the upper freshwater-marine transition, see discussion below; Fig. 15b). All transitions between the freshwater and marine diatom zones are diatomologically very distinct, implying a near total replacement of the assemblages. The assemblages in the upper freshwater zone, which in Vambheim-119 and Bu-113 occur in the gyttja above the marine diatom-bearing sediments (i.e., above 565 and 373 cm, respectively; Fig. 15a, b), are practically identical with the assemblages in the minerogenic sediments below these (i.e., in the depth intervals 569-612 cm and 407-420 cm, respectively; Fig. 15a, b). The upper freshwater zone above the organogenic-minerogenic sediment interface developed after the final isolation of these basins and thus undoubtedly reflects a lacustrine environment. Following the arguments of Mangerud (2000) we find it therefore quite remarkable that the lower freshwater zone, having a clear species compositional similarity with the upper zone, in reality should reflect a distinctly different (i.e., a marine or glacimarine) sedimentary environment.

ii) The diatoms in the freshwater zone below the marine zone in Vambheim-119 and in Bu-113 are in general well preserved (Fig. 16). The high degree of preservation of the diatom
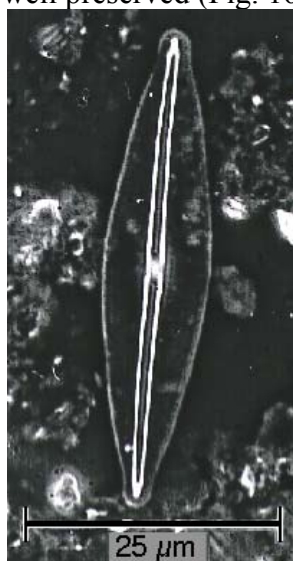
Fig. 16. A well-preserved specimen of *Frustulia cf.rhomboides v. saxonica* (Rabh.) de Toni from the lower freshwater diatom zone in Vambheim-119 (at 572 cm`s depth; Fig. 15a).

frustules suggests low influx of allochthonous valves into the basins and indicates that the freshwater diatoms were not transported by turbulent meltwater. In fact, the degree of preservation of diatom frustules in Vambheim-119 and Bu-113 is significantly higher in the freshwater-diatom bearing sediments compared to the marine-diatom bearing sediments. Since mechanical breakage and chemical corrosion of diatom frustules are generally regarded as stronger in marine environments than lacustrine environments (an important factor in this respect is that the rate of diatom dissolution generally increases in alkaline environments; see e.g., Lewin, 1961; Meriläinen, 1973; Flower, 1993; Barker et al., 1994), this adds support to the notion that the freshwater assemblages display lacustrine conditions and that the diatoms in the freshwater zones have in general not been exposed to sea water.

iii) In support of his view, Mangerud (2000) refers to the work of Jiang et al. (1998) who found 60-88% freshwater diatoms in sediment cores from the Kattegat, between Sweden and Denmark, interpreted as a result of meltwater influx from the Baltic Ice Lake. However, the freshwater diatom assemblages of Jiang et al. (1998) consist of a mixture of planktonic and non-planktonic forms. This is in contrast to Helle et al.'s (1997) material where the freshwater diatom assemblages are largely dominated by non-planktonic forms and where very few true planktonic freshwater forms occur. Planktonic forms are more frequently subjected to transport compared to non-planktonic forms (Simonsen, 1969).

iv) In Vambheim-119 and Bu-113 the fluvial input from rivers discharging directly into the basins was probably insignificant. (At present the drainage into these basins is mainly by seepage). Even more important is the fact that no large lakes exist in the catchments (or in the nearest upland areas) of Vambheim-119 and Bu-113 which in the past potentially could have supplied these basins with significant amounts of allochthonous freshwater diatoms. This favours the view of Helle et al. (1997) that the freshwater diatom assemblages in these basins are mainly dominated by autochthonous (intra-basinal) species.

v) The freshwater diatom assemblages of Helle et al. (1997) consist generally of more than 50% halophobous forms (i.e., forms which are strongly averse to sea water) and typically reflect acid, oligotrophic lake conditions. These are conditions to be expected in lakes in an arctic-like environment and with crystalline bedrock in their catchments.

vi) The environmental changes, as inferred from the diatom stratigraphy, are also generally reflected in the other microfossils in the sediment cores. Albeit in low concentrations, the numbers of marine dinoflagellate cysts are generally higher in the marine diatom-bearing sediment layers than in the freshwater diatom-bearing layers. In the marine diatom-bearing sediment layer in Vambheim-119 there is in addition a distinct peak of *Pinus* pollen (reaching more than 80 % of the total pollen sum), which is probably due to marine overrepresentation (see e.g., Florin, 1945).

vii) A comparison of diatom records with basin localities from other sites, open coastal (Sotra), as well as fjord-head sites (Yrkje), shows striking similarities (Fig. 15; see Fig. 1, for location). The compared basins are similarly situated with respect to the local marine limit (the upper basins are placed to the left in Fig. 15). The marine and lacustrine facies, as documented by the diatom records from basins at the outer coast, are analogous to the diatom records of Bu-113 and Vambheim-119.

viii)In Bu-113, the lower freshwater diatom-bearing sediment layer, interpreted as lacustrine, consists of coarse sand with gravel (depth interval 413-421 cm; Fig. 15b). The coarse-grained nature of this layer was taken by Mangerud (2000) as support for his interpretation of this layer as being marine or glacimarine. However, in Arctic or arctic-like environments coarse-grained (i.e., sand-rich) lake sediments are not uncommon (including basins with an insignificant fluvial input; cf. the palaeo-lake sediments of e.g., Vorren, 1978; Austad & Erichsen, 1987, pp. 35-38). In my opinion, it is more likely that the coarse-grained, freshwater diatom-bearing sediments in Bu-113 were derived from the littoral areas of the palaeolake. As previously noted, no major rivers were discharging directly into this basin. The basin is characterized by a deep, trough-shaped depression (the main coring site) surrounded by extensive shoal areas at near basin-threshold level (Fig. 17). When the sea level was below the basin threshold during the Lateglacial, large parts of the relatively flat promontory on which the basin is situated were subaerially exposed and the palaeolake likely experienced a complete seasonal lake ice cover (Fig. 17) (and probably also periods of extensive summer lake ice cover, as could be inferred from the low abundance of diatoms in the sand layer). Relatively large areas of the lake bed, in some places extending several hundreds of metres

out from the shore, were thus probably frozen to the ice cover. Littoral lake sediments (as well as littoral freshwater diatoms) frozen to the lake ice cover may have been deposited across the basin as the ice melted and broke up (cf. the various lake ice-related processes described by e.g., Shilts & Dean, 1975; Gilbert, 1990; Smith, 2000, based on studies of Arctic lakes). The basin morphology of Vambheim-119 is similar to that of Bu-113 (i.e., a deep trough-shaped depression surrounded by relatively extensive shoals, representing potential areas of seasonal freezing of the lake bed). Also in this basin, within the basal succession of sediments interpreted as lacustrine, there are sand and gravel-dominated layers (e.g., in the depth intervals 582-588 cm, 591-610 cm and 616-628 cm; Fig. 15a) interbedded with more fine-grained (silt-dominated) layers. Lacustrine ice rafting may also here provide a simple explanation of the origin of these coarse-grained, freshwater diatom-bearing sediment layers. In contrast, the situation at the head of Hardangerfjorden after the final isolation in the early Preboreal was quite different, implying non-arctic conditions with a denser vegetation cover in the catchments and less extensive lake ice cover, which in sum probably resulted in a strongly reduced influx of minerogenic sediments into the basins (and thus also in the transfer of coarse-grained littoral sediments to the central parts of the basins).
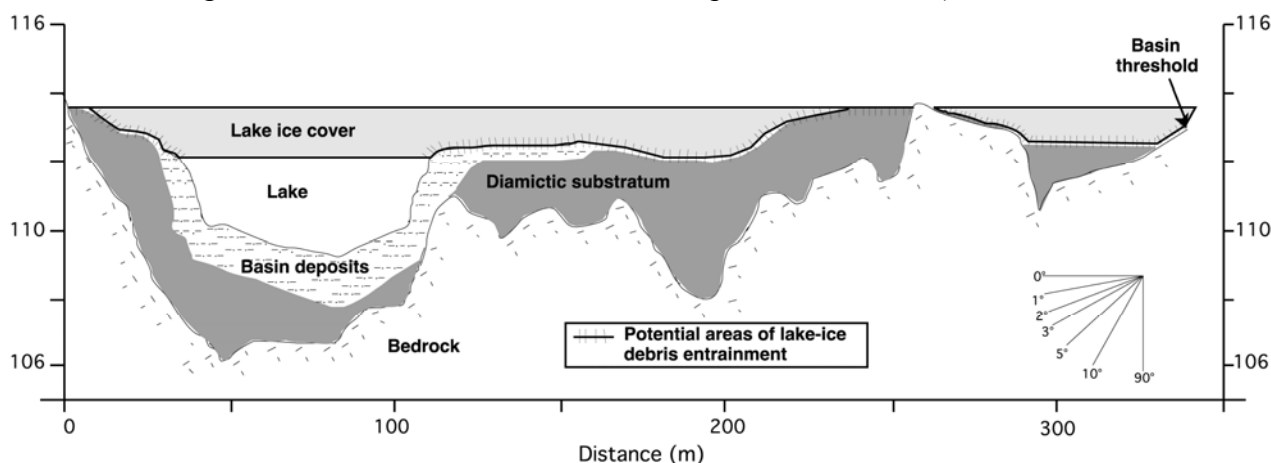

Fig. 17. A schematic section of the palaeo-lake Bu-113, showing a situation with a 1.5 m thick (seasonal) ice cover and where most of the littoral lake sediments are frozen to the ice cover. Vertical exaggeration 10.1:1. Modified from Helle et al. (1997).

Altogether these observations suggest that the succession of marine and freshwater diatom-bearing sediments in the lower part of the stratigraphy is a record of RSL change, as originally interpreted by Helle et al. (1997), and do not reflect fluctuations in the upland freshwater discharge and in the salinity of the surface waters of the fjord, as proposed by Mangerud (2000). A similar biostratigraphic pattern is also seen in the upper basins in Ljones and in Gravdal, in the mid parts of the fjord (i.e., in the basins situated less than ca. 15 m below the marine limit). In the lower minerogenic part of the stratigraphy in these basins, intervals with freshwater and/or brackish water forms (phytoplanktons and aquatics) occur below intervals with more salt-demanding brackish water and marine forms (see Paper 3 and Fig. 11). In contrast, in basins situated more than ca. 15 m below the marine limit, only a simple marine-to-freshwater succession (i.e., only a single isolation contact) is found.

However, in many basins (but not all!), both the pollen and the diatom sums in the lower, minerogenic part of the stratigraphy lie well below 300 throughout the interval. In these basins, the error limits for the estimated pollen and diatom percentages are large. For example, in the lower freshwater zone in Bu-113 the highest diatom sum obtained in one spectrum is 127 (at 420 cms depth, Fig. 15b). The probable error in percent at 95% level of confidence is estimated to between 1.5 and 7.5 %, depending on the species` relative occurrence in the spectrum (cf. the probable error monogram of Galehouse, 1971). For the neighbouring spectra, where the diatom sums are much lower, the error range is considerably higher and the assemblages found may not be representative

for this stratigraphic level. Clearly, this complicates the interpretation of the palaeoenvironment (see also Paper 3). However, in the other basin nearby, Vambheim-119, the diatom sums are mostly above 300 in the lower freshwater zone, implying that the observed shifts in diatom assemblages cannot be explained simply as a statistical artefact. In addition, some of the uncertainty has been reduced by analysing samples at densely spaced intervals, thus verifying that the neighbouring spectra show consistent, confirming trends.

Nevertheless, I cannot completely rule out the possibility that mechanisms other than RSL changes (e.g., salinity changes in the fjord; cf. Mangerud, 2000) may have been the dominant controlling factor on the diatom (and other microfossil) assemblages. This is the main reason why we started to investigate the Hæreid moraine (Paper 2). The aim of this investigation was to obtain a sea-level record independent of the isolation basin records, and thus, by using an approach totally different from that used by Helle et al. (1997), to test the hypothesis of an early post-deglaciation RSL rise at the head of the fjord.

6.4. The stratigraphy of the Hæreid moraine is incorrectly interpreted and cannot be used to estimate sea level.

In Paper 2, the conclusion that the Hæreid moraine underwent a major RSL cycle is drawn on the basis of the following observations and interpretations: A major palaeochannel cut deeply into the top of an ice-contact/proglacial delta, and which can be traced for 1 km along the NE side of the valley, is documented. It is clear that the channel incision took place by a subaerial river (and not by a subglacial river). The bedrock side-wall of the valley acted as a conduit for the channel flow, which started at the proximal side of the moraine and terminated at the distal, seaward side, where it entered the fjord on the upper parts of the delta's frontal slope. Later, the channel-fill/mouth-bar system was fed by a subaerial river in the same channel. The channel incision was most likely related to a distinct increase in the vertical accommodation space controlled by a lowering of the sea level. The base of the incised channel at the downstream reaches probably represents the time of maximum sea-level lowstand.

A major aggradation of the channel floor took place in the period that followed, as documented by the more than 10 m thick channel-fill deposits in the downstream reaches. Further downstream (i.e., on the palaeodelta's former frontal slope and prodelta zone), the GPR profiles show clinoforms with depositional dips directed outwards from the mouth of the palaeochannel. These clinoforms form part of the same (river-fed) clinoform system and are interpreted as mouth-bar foresets. The topset/foreset contacts can here be seen as progressively climbing in a seaward direction until they change to become vertically aggrading at the distal part of the complex, shortly before they reach the level of the surface interpreted as a wave ravinement surface (GPR profiles C-C'and E-E'; Fig. 5 in Paper 2). This indicates that, after the maximum sea-level lowstand was reached, the shoreline trajectory advanced basinward in an oblique-upward direction and thereafter changed to rise vertically. The vertical rise of the shoreline trajectory reflected by these deposits denotes the regressive–transgressive turnaround point of the depositional system and signifies the end of the lowstand phase.

The three near horizontal erosional surfaces seen on the northern flank of the channel-fill/mouth-bar complex (cf. the NW segment of profile A-A'; Sections 4.1 and 6.1 in Paper 2) are probably closely related to sea level. The lowest surface (at the base of unit H) occurs at approximately the same level as the lowest topset/foreset contacts identifed in the other channel-proximal parts of this system (interpreted as the maximum lowstand depositional shoreline break; profile C-C'). The next surface (base of unit J, interpreted as a transgressive surface) is approximately in level with the topset/foreset contact in the channel-distal parts of the system, at the point where the foreset beds start to aggrade vertically (see 'distal limit of shoreline progradation' in profile E-E'). This suggests that the base of unit J (downlap surface) reflects an increase in the water depth. The upper surface (the base of unit K, interpreted as a wave ravinement

surface) is approximately in level with the erosional surface that truncates the fan and delta foreset (also interpreted as a wave ravinement surface, profile B-B') and was formed after the shoreline had started to move landward.

As noted in Section 5.3 in Paper 2 there was, during the development of the channel-fill/mouth-bar complex, less accommodation space available in the area directly beyond the tip of the protruding bedrock wall (the area intersected by the NW segment of profile A-A') than in the other parts of the sedimentary basin outside the channel mouth. This explains probably why the mouth-bar units seen in the NW segment of profile A-A' are erosive-based, and why the foresets here (unit G and J) are vertically-stacked and significantly thinner than in the other parts of the palaeobasin. The near horizontal erosional surface interpreted as a wave ravinement surface has a possible extent of ca. 1.5 km. If most of the up to 8-m-thick succession above the wave ravinement surface is represented by transgressive shoreface deposits, then the sea-level rise during the transgressive phase was at least 8–10 m. These deposits are conformably overlain by laminated clay, interpreted as the remains of a highstand mud drape deposited as the rising RSL ultimately led to drowning of the moraine. Hence, according to this interpretation, the results from the study of the Hæreid moraine are compatible with the diatom-inferred sea-level changes.

Also observations from the neighbouring tributary fjords support the general picture of a regressive-to-transgressive marine shoreline development after the marine moraines here initially had aggraded to sea level. In Odda (Fig. 1), marine shells are found at the upvalley end of the basin, in deposits near the present-day shore of the moraine-dammed lake (Eitrheim, pers. comm. in Kvistad, 1965, p. 32), clearly demonstrating that the sea inundated major parts of the moraine here and invaded the basin proximal to this after the ice front had withdrawn from the area. GPR surveys of the moraine indicate that the basal deposits attributed to the ice-contact system are incised by a major (>10 m deep), infilled fluvial channel (Nygaard, 2002, p. 34). In the moraine's proximal parts, large-scale landward-dipping clinoforms occur, probably representing washover/floodtidal delta deposits corresponding to those found on the Hæreid moraine's proximal slope. On the top of the moraine in Odda, there is an up to 6 m thick subhorizontally layered unit (probably corresponding to the transgressive unit K in the Hæreid moraine) that caps the channel-fill and the washover/floodtidal delta deposits. Altogether, this indicates that the moraine in Odda was fluvially incised (initiated by a RSL fall) after the moraine had aggraded to the sea level, and that major parts of the moraine thereafter were inundated again and subject to transgressive ravinement.

However, the stratigraphy of the Hæreid moraine possibly allows for alternative interpretations. Below, I have presented and discussed some remarks on the aforementioned interpretations:

Remark 1: It is difficult to see arguments for some of the boundary surfaces indicated in the GPR profiles.

Reply: The sequence boundaries and the other boundary surfaces cannot be followed continuously in all segments of the GPR profiles. Some of the boundaries are, as also noted in the paper, drawn on a tentative basis and are thus more uncertain than others. It is for example very difficult to trace the wave ravinement surface in areas where it is not capping steeply dipping clinoforms, as in the area between the truncated delta top and the northern flank of the channel-fill/mouth bar complex. Here, the ravinement surface is reconstructed mainly by extrapolating from the northern flank area to the delta top (horizontal, dashed lines in profiles B-B', D-D' and E-E'), assuming that the surface extends across the entire area at about the same level. In some cases, the boundary surface between two different radar facies is somewhat masked by reflectors that merge and partly overlap. The boundary surface is thus located within a transitional zone and can be determined within given range of uncertainty. One such example is the palaeodelta front, which is determined on the basis of four GPR profiles. In all four profiles I have used the "last" NW dipping clinoform (delta foreset

bed) as an indicator of the frontal slope (profiles B-B' and D-D'). There is a marked change in the radar facies configuration seawards of the front (clinoforms dipping at 180° to the palaeodelta foreset), but the change occurs within an interval of a few tens of metres where the reflectors merge and partly overlap. Although the exact location of the frontal slope may be a matter of debate it is unlikely that it is located more than a few tens of metres updip or downdip of where I have indicated.

Remark 2: Could the surface interpreted as a wave ravinement surface instead represent an erosive topset base underlying preserved relicts of horizontally bedded Gilbert-type topsets?

Reply: The height equivalent, near horizontal erosional surface that truncates the top of the mouth bar succession at the distal end of the moraine (cf. Sections 4.1 and 4.3 in Paper 2) is probably part of the same (ravinement) surface. Hence, this surface is not only stratigraphically above the fan and delta foresets, but also stratigraphically above the channel-fill/mouth-bar complex. This is further substantiated by the observation that the topset/foreset contacts in the mouth-bar succession are seen in the profiles C-C'and E-E' as progressively climbing up to the level of this erosional surface. These foreset beds were formed after the shoreline had reached its maximum seaward position and downlap the transgressive surface (profile E–E`, Fig. 5 in Paper 2). In other words, the near horizontal erosional surface was formed after the distal limit of shoreline progradation was reached. This is difficult to reconcile with an interpretation of this surface as an erosive topset base. Moreover, the shallow wave base in the area, probably not deeper than 1.5–2 m, renders it unlikely that the deposits directly above this surface, up to 8 m thick, constitute the delta's topset, representing, e.g., braided-river alluvium and/or beach deposits. The deposit`s external, upvalley-thinning shape, and its internal reflection geometry (suggesting abundant onlap terminations; profile B-B`), are thought to be the result of a net upvalley propagation of the shoreline. Finally, evidence from the borehole data and the outcrop section on the moraine's proximal slope (cf. Sections 4.4 and 4.5 in Paper 2) suggest that the transgressive shoreface passed across the moraine's top surface and points to the interpretation of this surface as a ravinement surface.

Remark 3: Could the unconformity between units I and II in outcrop section 1 have been formed by glacial erosion or by subglacial meltwater?

Reply: The steeply inclined erosional surface between units I and II is located more than 70 m below the present top surface of the moraine and probably continues well below the base of the outcrop section. It must have formed far below the contemporaneous base level and cannot therefore be the result of erosion by a subaerial river. It is also unlikely that the unconformity was formed subglacially, by meltwater or glacial erosion. The absence of mud-rich diamictic sediments and the lack of any signs of glaciotectonic deformation in units I and II suggest that unit II is not subglacially deposited and do not form part of an ice-contact glaciomarine system. Instead, the lithofacies and the stratigraphic architecture exposed in the outcrop section are indicative of waterlain deposits. This is supported by the borehole data (Section 4.5 in Paper 2) which suggest that these sediments lie stratigraphically above the sediments of the ice-contact glaciomarine system. The unconformity is attributed to a major collapse of the moraine's proximal slope. Unit II above the unconformity is probably debrisflow deposits and the clinoforms of this unit are interpreted as the foreset beds of a flood-tidal delta. It is likely that the moraine's proximal slope experienced several failures during the transgression, particularly when the tidal delta built into the lagoon. On the other hand, the absence of till and glaciotectonic structures does not provide unambiguous proof that a glacier did not overrun the site. Hence, although the stratigraphic evidence makes this seem unlikely, I cannot completely rule out the possibility that the unconformity was formed by glacial erosion or by subglacial meltwater.

Remark 4: Major parts of the present-day surface are covered by a gravelly lag, including a large number of rounded boulders. Could this be remnants of a wave washed till?

Reply: Shallow borings conducted on the moraine's NE flank, near the NW segment of the GPR-profile line A–A`, show mainly fine sand below the topsoil and, nearer the terrace brink in the W, gravelly sand (Sognnes, 1982). Shallow trenches dug near the SE segment of the GPR-profile line B–B` show sandy gravel below the topsoil (Sognnes, 1982). The uppermost sedimentary unit in outcrop section 1 (unit V in the moraine`s proximal part), a 0–0.8-m-thick, erosive-based bed of sorted sandy pebbles and cobbles, is part of the same surface cover. The surface deposits are assigned to Lønne's (1995, 2001) allostratigraphic unit E, formed during the forced regression. Most likely, the gravel lag was formed during emergence, when the surface was exposed to wave action, and parts of the highstand mud and the underlying transgressive deposits were eroded. The boulder accumulations may also partly reflect the work of sea ice. The lag thus probably corresponds to the boulder lag found on the present-day marine beaches nearby. However, I cannot completely rule out the possibility that the lag is a remnant of a till, although I find it highly unlikely, given the evidence outlined above, including the stratigraphic evidence from the moraine`s upper, proximal part (see Remark 3).

Remark 5: Could the sedimentary succession revealed in the GPR profiles have been formed by an ice-front advance, implying a proglacial submarine build-up of ice-contact fan facies during the advance?

Reply: Above (under Remark 2), I argue why the near horizontal erosional surface that is capping the clinoforms of unit B and unit C (fan and delta foresets in profile B-B') and unit J (mouth-bar foreset in profile A-A') is interpreted as a wave ravinement surface (Section 4.2). The surface probably extends from the moraine`s ice-proximal, landward end to its ice-distal, seaward end at about the same level (but cannot be followed continuously in all segments of the GPR profiles, see under Remark 1) and probably developed mainly by shoreface retreat (sensu Fischer, 1961; Bruun, 1962; Swift, 1968; Nummedal and Swift, 1987; Demarest and Kraft, 1987). The channel-fill/mouth-bar system was fed by a subaerial river in the channel along the NE side of the valley. The channel starts at the proximal side of the moraine and terminates at the distal, seaward side. The steeply inclined clinoforms interpreted as mouth-bar foresets are part of the same (river-fed) clinoform system that has prograded outwards from the channel mouth. These foreset beds pass upstream into channel-fill deposits via a crestal levee/backslope zone. It is therefore highly unlikely that these deposits represent an ice-contact fan that was aggrading to sea surface. The lack of stratigraphic evidence for a late ice-front readvance in the moraine`s upper, proximal part supports this (see however under Remark 3). Instead, the development of the channel-fill/mouth-bar complex was probably closely related to the accommodation space created by a rise in sea level. Nearest the channel mouth, the foreset beds can be seen as prograding and aggrading up to the point where the lowest topset/foreset contacts in the complex are identified, occurring at an altitude of 97–98 m (profile C–C` in Paper 2). This altitude corresponds to the topset/foreset contacts between unit H and unit G in profile A–A`. The topset/foreset contacts in this complex (each corresponding to a depositional shoreline break, sensu Posamentier and Vail, 1988) can be traced continuously from their lowest levels, nearest the mouth of the palaeochannel, and basinwards, up to the level of the wave ravinement surface at an altitude of ca. 102 m (cf. Section 4.3 in Paper 2). The lowest topset/foreset contacts probably correspond to the level of the depositional shoreline break during the maximum lowstand. On this basis, I find it unlikely that the sedimentary succession, including any of the erosional surfaces [except for the deposits in the rear

zone of the moraine's submarine ice-contact fan (cf. unit A in GPR profile B–B`)], was formed by an advancing ice front.

Remark 6: Could the channel downcutting be explained by powerful meltwater events (jökulhlaups?), and the clinoforms outside the channel mouth represent the resultant deposits of such events? Given this setting, could the channel-fill/mouth-bar complex have developed purely under a forced regression (i.e., during a RSL fall)?

Reply: During a forced regression, the shoreline translates seaward and obliquely downward, independently of sediment supply (and of flow regime) (cf. Posamentier et al., 1992). High sediment supply (and flow regime) may result in high rates of delta progradation, but will not change the shoreline trajectory so that the shoreline builds seaward horizontally or obliquely upward (even in the case when the basin physiography allows basinward-shallowing conditions to develop; cf. Helland-Hansen & Martinsen, 1996). Most likely, the downcutting of the channel was initated by a RSL fall and the resultant deposits were in the form of foreset beds deposited outside the river mouth on the former delta face. As sea level was falling, the foreset beds were deposited at successively lower levels (downward-stepping clinoforms; cf. Phase 2; Fig. 10 in Paper 2). Deposition within the channel was limited. The downcuttng is likely to have been caused either by rivers draining an (ice-contact) proglacial lake dammed by the moraine or by meltwater rivers discharging directly from the ice front while a glacier still occupied the valley basin behind the moraine. In both instances, a high flow regime was probably maintained throughout the downcutting phase and the subsequent infilling phase. The development of the channel-fill/mouth-bar complex on the delta's former frontal slope and prodelta zone represents a basinward shift of depositional systems. As the RSL stabilized, the rate of vertical accommodation space generated at the shoreline diminished significantly (compared to the previous forced regressive phase). This forced the fluvial/deltaic depositional systems to migrate basinward. By this stage, the channel incision in the downstream end had mainly ceased. As the RSL began to rise, accommodation was also generated behind the shoreline, giving space for net aggradation of channel-fill deposits. The complex began to form by normal regression (and not by forced regression) as can be read from the topset/foreset contacts in this complex, which can be traced basinward in an oblique-upward direction.

Altogether, it is very difficult to explain the stratigraphy and depositional architecture of the Hæreid moraine (i.e., within the upper 15–20 m of the sedimentary succession) simply in terms of a RSL fall. Hence, if the alternative RSL curve given in Paper 3 (Fig. 18b) is correct, many of the observations made on the Hæreid moraine, and on the other marine moraines nearby, would in my opinion remain largely unexplained. However, the absolute age control is poor and, as already noted, the absence of till and glaciotectonic structures do not provide unambiguous proof that a glacier did not overrun the site. I cannot therefore completely rule out the possibility that the Hæreid moraine at some stage was ice overrun during the YD.

7. Conclusions

In the present thesis, the stratigraphical evidence of a major RSL rise correlated with the YD transgression (Papers 1-3) has been taken to indicate that Hardangerfjorden became deglaciated prior to the YD (i.e., in the Bølling/Allerød period) and that the fjord, or the main part of it, remained ice-free in the following period. This is in conflict with the published reconstructions of the YD ice margin, indicating an ice-filled fjord during this period. The implications of a scenario with an ice-free fjord for the interpretation of the stratigraphic and other glacial geological evidences in the area are discussed. Although there is a possibility that the shell-bearing diamictons in the outer Hardangerfjorden area may have a non-glacial (glaciomarine) origin and/or that some of the moraines may have been formed by a land-based ice, many observations [including those
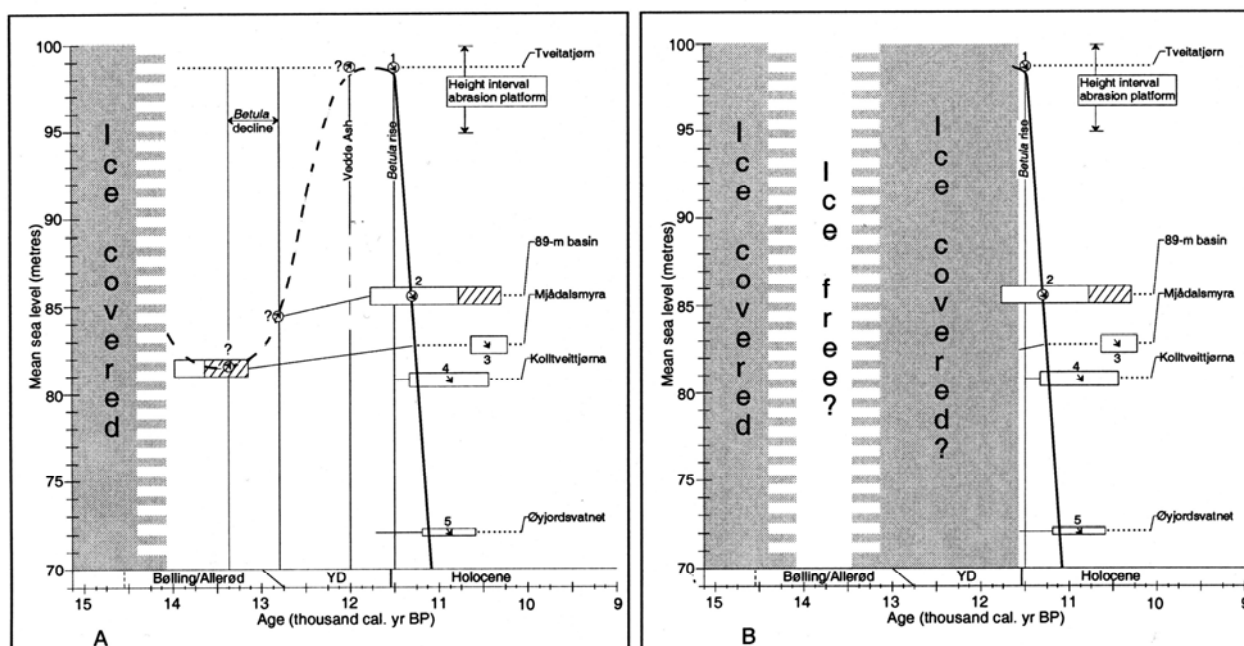
Fig. 18. Two alternative sea level curves for the Ljones area, mid Hardangerfjorden area. The curve to the left (A) accommodates better the observations from the mid and inner parts of the fjord, whereas the curve to the right (B) accommodates better the observations from the outer Hardangerfjorden area, indicative of an ice-filled fjord during the YD. Figure from Paper 3.

made by Follestad (1972) on the mainland north of Halsnøy and by Lohne et al. (2003) on Tysnes] remain unexplained in the context of an ice-free fjord. In seeking a solution this conflict, our interpretations of the sea-level records along Hardangerfjorden are critically evaluated, and alternative interpretations are discussed. The possibility that mechanisms other than RSL changes (e.g., salinity changes in the fjord; cf. Mangerud, 2000) may have been the dominant controlling factor on the diatom (and other microfossil) assemblages in the isolation basin deposits cannot be ruled out. However, when viewed together with the observations from the Hæreid moraine (and the other marine moraines in the area), RSL change is in my opinion the mechanism that best accounts for the observed biostratigraphic and sedimentary successions. The synthesis of glacial geological and shore level features from the Hardangerfjorden area suggests that the data cannot be placed into one all-encompassing glacial model. I therefore consider the deglaciation history of the fjord as currently still unresolved and that many questions are yet to be solved.

If we look beyond the debate on the fjord`s deglaciation and earliest sea level history, I consider the most important contributions of the present thesis to have been: (1) to highlight marine moraines as potentially highly sensitive recorders, not only of glacier dynamics, but also of sea-level changes, (2) to document the complexity of the stratigraphic architecture of such landforms, and (3) to demonstrate how the use of sequence stratigraphic principles may help to extract information from these sedimentary archives. Also of value is the documentation of marine limits along the fjord by means of stratigraphic investigations. These data add constrain on the timing of the onset of the early Holocene sea level regression and on the rapidity of the RSL fall. Finally, the compilation of Late Weichselian and early Holocene sea-level data from the region is considered to be useful, not only because the data are now made more available for the scientific community, but also because this work highlights the potential importance of neotectonic faulting in shoreline reconstructions.

8. Recommendations for future research

More sedimentary basins (lakes and peat bogs) located inland should be investigated, with emphasis on establishing an accurate chronology of the basal, early post-deglaciation deposits. If possible, this should include densely spaced AMS-dates of terrestrial plant remains.

A detailed reconstruction of the stratigraphic and sedimentary architecture of the Halsnøy moraine and adjacent deposits should get high priority. Before the internal organisation of the main depositional units is sufficiently known, it is difficult to see how the ice-contact system(s) in these landforms can be precisely dated. A detailed reconstruction of the stratigraphic and sedimentary architecture will undoubtedly also add more insight into the genesis of lithofacies such as the Grindavoll silt and the shell-bearing, till-like diamictons.

Although excellent allostratigraphic models already exist for marine moraines (e.g., Lønne, 2001), I feel in general that much work still remains to be done when it comes to our understanding of how these landforms evolved under the control of sea-level changes. (For example, Lønne's model are based on case studies that are mainly from areas where the sea-level changes are characterised by a monotonic fall over the period during which they have been ice-free.). More studies of such landforms are therefore needed.

The use of sequence stratigraphic methods in the analysis of Quaternary sedimentary successions is highly recommendable. Although there is a long tradition of sea-level research in Fennoscandia only a few analyses of Quaternary deposits have been undertaken using sequence stratigraphic methods, despite that the principles have been known for several decades (Jensen et al., 1997; Bennike et al., 2000; Leiknes, 2000; Jensen et al., 2002). The concepts of sequence stratigraphy are scale- and time-independent (see e.g., Posamentier et al., 1992) and the stratigraphic methods have proven to be a powerful tool for the analysis of sedimentary-basin evolution and the prediction of stratigraphic and sedimentary architecture [see however Winsemann et al. (2004) for a discussion of potential pitfalls when applying such principles in previously glaciated terrains]. Mapping of physical surfaces in the subsurface using ground-penetrating radar techniques provides a good basis for sequence stratigraphy. These techniques allow the observer to track stratigraphical horizons over long distances. Individual, localised lithofacies exposed in outcrop sections and in boreholes can in this way be seen in a wider context. Conversely, analyses of Quaternary deposits may also yield important conceptual information on high-resolution sequence stratigraphy.

At present there exists no satisfactory explanation of the cause(s) of the YD transgression. The theoretical model of Anundsen and Fjeldskaar (1983), where this RSL rise is interpreted to be the result of the interplay between glacio-isostasy (i.e., a flexural re-depression of the lithosphere) and geoidal eustasy (i.e., gravitational attraction between the ocean and the growing ice masses), superimposed on the rising trend of glacio-eustatic sea level, needs to be revised. A number of independent observations, including new values for the flexural rigidity of the region's lithosphere (cf. Djomani et al. 1999, Rohrman et al. 2002) and new sea-level data from the region, are not conformable with Anundsen and Fjeldskaar's (1983) theoretical predictions (see discussion in Paper 4). Future modelling should include all information that presently is available on the mantle and the lithosphere, as well as on the spatial and temporal components of the YD transgression. Ideally, the new model should contain a minimum of adjustable parameters (e.g., parameters that are dependent on the choice of ice model) and should be robust with respect to changes in those parameters.

Efforts should be made to identify and study faults in the region which potentially carry components of postglacial displacement. Only in exceptional cases are neotectonic features evident from surface exposures. Therefore, to examine an area for neotectonic activity, different techniques and approaches should be used. Detecting potential neotectonic faults is important, not only from an academical point of view, but also for the society and its infrastructure.

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