Seismic stratigraphy of Younger Dryas ice-marginal deposits in western Norwegian fjords

INGE AARSETH, PER KRISTIAN AUSTBØ & HARALD RISNES


The submarine portions of ice-marginal moraines and their distal sediments have been studied in the fjords between Stavanger and Åndalsnes, western Norway (59°–62°30'N). Echosounder, sparker, and high-resolution seismic profiles are used to record the thickness, morphology and structure of the deposits, which are divided into four main seismic stratigraphic sequences describing the different phases of sedimentation. Evidence of several ice-push events is observed in some of the fjords, and models of glacial dynamics are presented for deep water and shallow water environments. Volumes of the different sedimentary facies have been estimated for some of the fjords. Coarse sediments (mainly glaciofluvial foreset beds) represent less than 10% of the Younger Dryas sequence. Dating of the moraines is accomplished through morphological correlation to the established chronostratigraphy of terrestrial sections.

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Introduction

The mapping of Younger Dryas end moraines in western Norway has shown that the ice margin crossed many fjords (Fig. 1). Terrestrial mapping of the moraines was undertaken by a number of authors, from south to north (e.g. Andersen 1954; Undås 1963; Holtedahl 1967, 1975; Maisey 1968; Fareth 1970, 1987; Aarseth 1971; Anundsen 1972; Follestad 1972; Aarseth & Mangerud 1974; Kramer 1977; Sollid & Sørbel 1979; Aa 1985; Klakegg et al. 1989; Thoresen et al. 1995). Echosounder data were presented to provide additional information on the submarine topography of the supposed moraine thresholds (Aarseth 1971; Anundsen 1972; Aarseth & Mangerud 1974).

The suggested composition of these thresholds were nevertheless speculative until reflection seismic profiling provided evidence for their variety and complexity (Seland 1981; Giskeødegaard 1983; Aarseth 1988; Aarseth 1988; Austbø 1988; Risnes 1990). Mapping of this kind has demonstrated large ice-marginal accumulations in some of the fjords in Nordland and Trøndelag, 100–400 km northeast of Kristiansund (Fig. 1; Rokøngen 1979; Andersen et al. 1982; Bjerkli & Olsen 1990; Reit 1994; Ottesen, Frenstad & Rokøngen 1995).

Our investigations of the fjord sediments in western Norway have been focused on the internal structure of the moraines and their distal sediments, and their relation to the general glacial geology of the area.

Sedimentary environment at the margins of fjord glaciers

Sedimentary environment at the margins of fjord glaciers are described from modern tidewater glaciers (Powell & Molnia 1989; Syvitski 1989), and several models for such environments are put forward (Powell 1981, 1984; Elverhøi, Lønne & Seland 1983; Syvitski, Burrel & Skei 1987). General models for submarine glacial deposition also describe sedimentation processes near the grounding line (Orheim & Elverhøi 1981). Andersen et al. (1981) describe a generalized cross-section of a sub-aequously deposited terminal moraine, and the depositional history of Younger Dryas ice-contact deltas are reconstructed on the basis of the interpretation of sedimentary facies (Brandal & Heder 1991; Lønne 1993, 1995). Morainal bank systems may consist of several different facies because of the variety of processes that contribute sediment at the grounding line (Powell & Molnia 1989). Long-term changes in mass balance leading to major glacier advances influence the sedimentary environment at the glacier terminus (Boulton 1986).

Younger Dryas glacier fluctuations and size of the moraines in western Norway

Considerable advance of the Younger Dryas ice front in western Norway was documented by 14C datings of molluscs in sub-till marine sediments as well as of shell fragments in till (Undås 1963; Holtedahl 1964, 1967, 1975; Mangerud 1970; Anundsen 1972; Follestad 1972; Aarseth & Mangerud 1974; Fareth 1970, 1987; Mangerud et al. 1979; Rye et al. 1987). In spite of the many dates both from proximal and distal sites, the readvance itself is not very accurately dated. An age closer to 10,500 BP than 10,000 BP is considered most likely (Mangerud 1980). 14C dates are listed in Table 3.

The terrestrial marginal deposits vary from generally small ridges 1–5 m high in the Bergen area (Aarseth...
In some areas 14C dated glacial tectonized sediments have led to suggestions of an early Younger Dryas ice advance several km beyond the known terminal moraines, yet without any morphological evidence. Anundsen (1977) and Blystad & Anundsen (1983) report this in the Boknfjord area, and Sindre (1980) has observed similar evidence on the Stord island. In the Nordfjord area lateral moraines beyond the main Younger Dryas moraines (Fareth 1970), suggest a marginal position 13–15 km further west (Fareth 1987). 14C dates have since confirmed an early Younger Dryas age for this event, but terminal moraines have so far not been found (Klakegg & Nordahl-Olsen 1985). Giskeødegaard (1983), has suggested ice-advances beyond the marginal moraines in the Sunnylvsfjord and the Norddalsfjord (Fig. 1) based on interpreted seismic profiles.

Methods

Fieldwork in the fjords was carried out from two research vessels: ‘Hans Reusch’, and ‘Håkon Mosby’, with the same seismic equipment (EG&G sparker and boomer), but with different echosounders. Calculations of surface slope angles are based on the higher frequency data. Detailed bathymetry of the fjord basins was obtained from hydrographic originals (1:20 000) made available by the Norwegian Hydrographic Office.

To penetrate the thick sediment units in the deeper fjords as well as the coarse sediments of the moraines, a 1 kJ, 3-element sparker was used. The analog signals were received through a single channel streamer with thirty hydrophones for the ‘Håkon Mosby’ (150 feet), and eight hydrophones for the smaller ‘Hans Reusch’ (34 feet). Navigation on both ships was carried out as a combination of radar and dead reckoning. A combination of transverse and longitudinal profiles in all areas except for the narrowest straits served to minimize side-echo effects as well as to provide a three-dimensional picture of the sediment sequences in the steep-sided fjords. Profile grid density varies and is listed in Table 1.

<table>
<thead>
<tr>
<th>Fjord area</th>
<th>Area (km²)</th>
<th>Profiles</th>
<th>Profile (km)</th>
<th>Length of fjord basin (km)</th>
<th>Year profiled</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jøsensefjord</td>
<td>23</td>
<td>26</td>
<td>72</td>
<td>26</td>
<td>1984–85</td>
</tr>
<tr>
<td>Ernfjord</td>
<td>19</td>
<td>12</td>
<td>23</td>
<td>11</td>
<td>1984–85</td>
</tr>
<tr>
<td>Sandsfjord</td>
<td>33</td>
<td>22</td>
<td>53</td>
<td>×</td>
<td>1984–85</td>
</tr>
<tr>
<td>Vindafjord</td>
<td>37</td>
<td>16</td>
<td>37</td>
<td>30</td>
<td>1984–85</td>
</tr>
<tr>
<td>Herdla area</td>
<td>1</td>
<td>16</td>
<td>8</td>
<td>×</td>
<td>1985</td>
</tr>
<tr>
<td>Fenessfjord</td>
<td>12</td>
<td>11</td>
<td>22</td>
<td>40</td>
<td>1986</td>
</tr>
<tr>
<td>Sandøy area</td>
<td>×</td>
<td>16</td>
<td>20</td>
<td>×</td>
<td>1986</td>
</tr>
<tr>
<td>Sognefjord</td>
<td>×</td>
<td>200</td>
<td>500</td>
<td>200</td>
<td>1979–86</td>
</tr>
<tr>
<td>Dalsfjord</td>
<td>20</td>
<td>28</td>
<td>45</td>
<td>55</td>
<td>1986–87</td>
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<tr>
<td>Førdefjord</td>
<td>10</td>
<td>22</td>
<td>27</td>
<td>70</td>
<td>1986–87</td>
</tr>
<tr>
<td>Hundvikfjord</td>
<td>12</td>
<td>10</td>
<td>23</td>
<td>110</td>
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<tr>
<td>Sunnylvsfjord</td>
<td>17</td>
<td>23</td>
<td>42</td>
<td>40</td>
<td>1984</td>
</tr>
<tr>
<td>Norddalsfjord</td>
<td>15</td>
<td>18</td>
<td>36</td>
<td>25</td>
<td>1984</td>
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Table 2. Sediment volumes in the three seismic sequences in some of the investigated fjord areas.

<table>
<thead>
<tr>
<th>Fjord area</th>
<th>Area (km²)</th>
<th>Volume (km³)</th>
<th>Seq. 1 (km³)</th>
<th>Seq. 2 (km³)</th>
<th>Seq. 2: Coarse part (km³)</th>
<th>Seq. 3 (km³)</th>
<th>Total in fjord (km³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jøsenfjord</td>
<td>23</td>
<td>0.84</td>
<td>0.36</td>
<td>0.47</td>
<td>0.08</td>
<td>0.01</td>
<td>x</td>
</tr>
<tr>
<td>Erfjord</td>
<td>19</td>
<td>0.23</td>
<td>x</td>
<td>x</td>
<td>0.04</td>
<td>0</td>
<td>x</td>
</tr>
<tr>
<td>Sandsfjord</td>
<td>33</td>
<td>0.34</td>
<td>0.16</td>
<td>0.36</td>
<td>0.10</td>
<td>0.04</td>
<td>19</td>
</tr>
<tr>
<td>Vindafjord</td>
<td>37</td>
<td>0.23</td>
<td>0.12</td>
<td>x</td>
<td>0.02</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>Fensfjord</td>
<td>12</td>
<td>0.70</td>
<td>0.30</td>
<td>0.36</td>
<td>0.10</td>
<td>0.04</td>
<td>19</td>
</tr>
<tr>
<td>Sognefjord</td>
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<td>x</td>
<td>0.04</td>
<td>0</td>
<td>x</td>
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</tr>
<tr>
<td>Sundafjord</td>
<td>20</td>
<td>0.80</td>
<td>0.55</td>
<td>0.20</td>
<td>0.01</td>
<td>0.05</td>
<td>3</td>
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<td>Førdefjord</td>
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<td>0.90</td>
<td>0.40</td>
<td>0.35</td>
<td>0.10</td>
<td>0.02</td>
<td>12</td>
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<tr>
<td>Hundvikfjord</td>
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<td>0.60</td>
<td>0.23</td>
<td>0.35</td>
<td>0.10</td>
<td>0</td>
<td>7</td>
</tr>
<tr>
<td>Sunnylvsfjord</td>
<td>17</td>
<td>1.30</td>
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<td>0.80</td>
<td>0.07</td>
<td>0</td>
<td>x</td>
</tr>
<tr>
<td>Norddalsfjord</td>
<td>15</td>
<td>1.00</td>
<td>0.35</td>
<td>0.55</td>
<td>0.04</td>
<td>0</td>
<td>x</td>
</tr>
</tbody>
</table>

Ordinary seismostratigraphical principles were used in the interpretation of the profiles. Seismic reflectors vary in amplitude, frequency and continuity (Leenhardt 1971), and are the result of changes in acoustic impedance (Leenhardt 1969). Seismic sequences can be established on the basis of uniform reflectors, and the reflectors can generally be considered as time-stratigraphical horizons rather than lithological boundaries (Nystuen 1989). Lateral variations within one sequence are thought to represent changes in facies and can aid in the interpretation of the lithology (Boulton, Chroston & Jarvis 1981). Near-horizontal basin-floor reflectors in distal glaciomarine sediments are shown to represent the rhythmic input of sediment loaded meltwater (Aarseth, Lønne & Giskegaard 1989). Such reflectors can also develop as a result of compaction during periods of non-deposition (Leenhardt 1971) or represent changes in grain size. Sediment thicknesses are measured in ms TWT (milliseconds two-way time) and converted into m by a factor of 0.8 using a sound velocity of 1600 m/s for the total sequence.

The ice-marginal deposits are described from south to north (Fig. 1) with representative examples of raw and/or interpreted profiles. In some areas the glacial history, as gathered from studies of terrestrial sections, is enhanced by correlation with the marine deposits. The deposits are divided into four seismostratigraphic sequences encompassing different phases of sedimentation: Sequence 1 is deposited during the Allerød deglaciation. Sequence 2 is deposited at the quasi-stable ice front during the Younger Dryas, and Sequence 3 is deposited during possible succeeding glacial advances. Sequence 4 represents later redeposition by slides or deposition of younger glaciomarine and Holocene sediments. Isopach maps are provided for some of the areas and sediment volumes are presented in Table 2.

Areal descriptions: Boknfjord area

The Boknfjord is an open embayment with several islands and with narrow and deep tributary fjords cut into the landmass (Figs. 1, 2). This embayment, and the Jæren low-land to the south, were the earliest deglaciated areas in southern Norway and became ice free at about 13,000–13,500 BP (Andersen, Wangen & Østmo 1987) or possibly as early as 14,000 BP (Anundsen 1985; Paas 1990). The inner parts of Boknfjord were deglaciated just after 12,000 BP. At Hjelmeland (Fig. 4), just outside the mouth of Jøsenfjord, a dated section points to an early Younger Dryas ice advance, possibly caused by a surging glacier (Bløstad & Anundsen 1983). The position of the Younger Dryas moraine is very close to the mouth of the tributary fjords (Fig. 2).

Lysefjord

The terrestrial parts of the Younger Dryas moraine at Lysefjord (Fig. 1) are prominent ridges and outwash...
terraces on both sides of the fjord entrance (Andersen 1954). A few km to the east of the mouth lies the moraine known as the 'Esmark moraine' where the classical observation that led to the theory of continental glaciation was made by Esmark in 1824 (Andersen & Borns 1994).

Because of the shallow morainic threshold (13 m) multiples as well as the use of the long hydrophone cable have obscured the internal reflectors, but the acoustic basement reflector is most probably the crystalline basement revealing a maximum sediment thickness of ca. 200 m. The distal slope into the Høgsfjord is a maximum of 16° (Fig. 3).

Jøsenfjord

The 25 km long, 1–1.5 km wide and 600 m deep Jøsenfjord terminates at a 145 m deep sill (Fig. 4A) just west of the two bedrock points containing morainic spurs (Fig. 2). A terrace at Mula defines the Late Glacial marine limit of 62 m a.s.l. in this area, and the nearly vertical fjord sides confined the 1300 m thick Younger Dryas fjord glacier (Andersen 1954).

Depths to acoustic basement (Fig. 4B) clearly demonstrate the continuation of the sub-sediment Jøsenfjord into the Gardsundfjord with only a minor bedrock sill in between. Sediment thickness at the sill is 300 m, decreasing to 200 m in the distal basin (Fig. 4C).

Fig. 5A shows one of the sparker profiles along the fjord axes and Fig. 5B shows the 3-sequence seismostratigraphic breakdown. In the distal part of the section the
lower part of Sequence 1 has acoustically even, parallel lamination filling the basins (onlap). Core samples of units with similar acoustic configuration represent distal glaciomarine silty clays sometimes containing thin silt and fine sand laminae (Aarseth, Lønne & Giskeødegaard 1989). The upper part of Sequence 1 is wavy and somewhat contorted in the distal part and discontinuous with a lower amplitude in the proximal part. This is thought to be caused by a more proximal glacial environment with possible glaciotectonic disturbances of the proximal sediments and shifting glacial drainage with sediment gravity flows and turbidites forming channels in the distal part. The lower amplitudes of the proximal part may also be caused by the steep and uneven surface slope (ca. 10°) with incident reflection of parts of the signals.

The bounding reflector to Sequence 2 displays a high amplitude under the sill and forms an angular unconformity in the distal part. The central part of Sequence 2 is a deltaic unit where the eastern foreset layers display an increasing inclination (3–9°) and a downlap on to the lower boundary. The increase in inclination is thought to be caused by a gradual transition from a squeezing of diamicton material at the grounding line to more sorted material brought forward by increasing meltwater activity. The youngest foreset layers of Sequence 2 are parallel (ca. 9° dip), and the upper part of the sequence has an angular unconformity overlain by a hummocky part and a nearly horizontal, high amplitude upper bounding reflector. The 2/3 boundary is not marked in the distal part due to lack of continuity and a more gradual transition.

On the longitudinal profile Sequence 3 reaches max. 20 m thickness. This nearly horizontal sequence caps the sill area and forms a small ridge near the distal crest. The distal foreset beds with lower inclinations are most probably parts of this sequence. On transverse sections (Fig. 6), Sequence 3 comprises lateral ridges up to 65 m thick leading to the lateral morainic spur of Porsberg and Skjerterangen (Fig. 4). This profile also shows small ridges and channels, max. 14 m deep and 120 m broad, leading up the proximal slope. Further down this slope lies a prominent sediment ridge in the middle of the fjord (Fig. 4A). The acoustic reflectors below the proximal slope are
more or less parallel to the slope (Fig. 5). They may be part of Sequence 3 or alternatively Sequence 2. In the latter case the ridge could represent a feeding esker to the deltaic unit as described from moraines in Lake Superior (Landmesser et al. 1982). Glaciofluvial material deposited up-slope has also been found in early Preboreal ice-contact deposits in eastern Norway (Holtedahl 1974) and the actual process has been observed in a fjord in Spitsbergen (Liestøl 1973).

Jøsenfjord moraine: Depositional history

Seismic sequences at the mouth of the Jøsenfjord suggest a complex history of deposition which conforms with the interpretation of the stratigraphy at Hjelmeland (Blystad & Anundsen 1983). Sequence 1 represents glaciomarine deposits from a glacier within the Jøsenfjord during the Allerød chronozone. The glacier then advanced over the sill and into the embayment of Hjelmeland between 11,100–10,800 BP. Glacial erosion formed the bounding reflector between Sequences 1 and 2 in the sill area and density currents, together with mass movements, probably made the acoustically contorted zones in the distal part.

A glacial terminus for this advance cannot be determined from the seismics alone, but according to Blystad & Anundsen (1983) the glacier did not reach Randøy. After a retreat to the steep proximal slope, a phase of glaciofluvial deposition formed the deltaic part of Sequence 2 where the increasing foreset slopes probably indicate an increasing sorting of the sediments. The slope-parallel reflectors down the proximal slope may be a feeding esker in this sequence.

Sequence 3 with its strong bounding reflector is thought to represent a final glacial advance on the sill causing some glaciotectonic disturbances of the upper part of Sequence 2. The ice reached the outer crest where it formed the western-most foreset beds and a small morainic ridge on top of them. The loading by the glacier at the crest of the moraine probably caused the slumping that steepened the terminal slope (ca. 17°) and formed the lobes at the base of the delta front. The large lateral ridges (Fig. 6) were probably formed during this phase, mainly from debris produced along the steep sides of the Jøsenfjord proper. The exact timing of the final glacial retreat is not known in the Jøsenfjord area (Blystad & Anundsen 1983).

Erfjord

The 1 km wide and 8 km long Erfjord terminates at the eastern end of the wider and deeper Jelsafjord (Fig. 2). The Younger Dryas terminal moraine is situated at the mouth of Erfjord, similar to the situation in the Jøsenfjord. Lateral moraines confine the extension of the glacier to the south and north (Lyngsnes and Landsnes respectively: Anundsen 1972), and the saddle point of the moraine is presently at 130 m water depth.
Fig. 8. Sparker profile 64–73 across the Sandsfjord 1.4 km proximal to the Younger Dryas moraine. For location, see Fig. 2.

representing tectonized glaciomarine sediments (Sequence 1). Younger undisturbed sediments (Sequence 4) are deposited in the channels on either side of the ridge.

This part of the Sandsfjord has a relatively low relief. A fjord glacier confined to the narrow and deep fjord channel further north became divergent and may have been less erosive, and therefore deposited the drumlinoid form on the lee side of the bedrock high. Between Foldøy and Hebes, a less prominent ridge crosses the relatively shallow (85 m) sound.

Vindafjord

The Younger Dryas glacier in the max. 850 m deep Vindafjord was fed from the Inner Sandsfjord over the Ropeid landbridge (Undås 1963; Anundsen 1972). At the intersection of Sandeidsfjord and Yrkjefjord the glacier split in three branches towards north, west and south. The terminal moraines in the Yrkjefjord and the Vatsfjord were mapped by the use of a conventional echo-sounder. Similar mapping near Askvik in the Vindafjord did not define any moraine-like ridges and a floating terminus was postulated in the 350–400 m deep fjord (Anundsen 1972).

A seismic grid between Askvik and Amdal unveiled a 75 m high and 100 m thick morainic spur southeast of Amdal (Austbø 1988). This asymmetric position corresponds to the location of the deep fjord channel situated on the western side of the fjord. The max. 120 m thick distal sediments terminate as several slide scars out into the 700 m deep Nedstrandsfjord (Fig. 2).

Boknfjord area: Summary

All the moraines crossing fjords in the Boknfjord area, with the exception of those in the Yrkjefjord and Vats-

fjord, are situated at the transition between the narrower tributary fjords and the wider and deeper main fjord (Fig. 2). The acoustic basement below the moraines is usually somewhat shallower than the rest of the fjord channel. The moraines, however, cannot be considered as recessional moraines as the seismic stratigraphy clearly shows one or more ice-push events. This is in good agreement with the stratigraphy established from terrestrial sites. Except for Sandsfjord, the areas proximal to the ridges have very thin sediment cover, and all areas show relatively thick units of distal sediments.

Hardangerfjord area

The Hardangerfjord is a southwest–northeast trending fjord mainly eroded along the border between Precambrian rocks on the south-eastern side and Caledonian rocks on the north-western side (Holtedahl 1967, 1975). The fjord has different names for each of the various parts (Fig. 9). The terrestrial Younger Dryas moraines are very prominent in this area. The main fjord glacier terminated between the islands of Huglo and Halsnøy at the threshold between the Husnesfjord and the Klosterfjord, and formed the moraine that constitutes parts of the Halsnøy (Fig. 9; Undås 1963; Holtedahl 1967, 1975; Aarsæth & Mangerud 1974). An eastern branch of this glacier reached Sandvoll in Høylandsundet. To the east, the glacier in the Skåneviksfjord was fed by glaciers in Åkrafjord and Mårefjord and terminated between Skånevik and Ølfarnes (Anundsen 1972; Folloestad 1972).

There are few 14C datings from this area. At Valen, just proximal to the moraine, shells in a clayey till date to 11,470 ± 180 BP (Holtedahl 1967). At Ølve, 25 km proximal to the moraine, shells in overconsolidated clay are dated to 11,230 ± 180 BP. This gives a minimum distance of glacial retreat in Allerød (Aarsæth & Mangerud 1974).

Fig. 9. The Skånevik–Halsnøy–Huglo area. Younger Dryas terrestrial moraines according to Folloestad (1972) and Holtedahl (1975). For location, see Fig. 1.
An ice advance to the island Stord just after 11,000 BP has been claimed on the basis of dates at Leirvik (Sindre 1980). Shells in glaciomarine sediments at Jektevik on the eastern side of the island were dated to 10,600 ± 100 BP (Genes 1978) and give a minimum date of the deglaciation of the Langen outside the moraine at Tysnes.

Skåneviksfjord

Fig. 10 shows an interpretation of a sparker profile along the northwestern part of the Skåneviksfjord. The ca. 100 m thick glaciomarine sediments (Sequence 1) are overlain by max. 80 m thick and gently dipping foreset beds (Sequence 2). On this side of the fjord channel no morainic material can be seen capping the foresets.

A seismic profile grid in the fjord has revealed an asymmetric location of the glaciofluvial foreset beds. In the middle and southeastern part only a low hummocky morainic ridge occurs. The glaciofluvial drainage has evidently been guided by the deep fjord channel and directed around the point at Vannes. The glacier from the Skånevik valley may have caused a northwestward slope of the glacier surface near the terminus, forcing the meltwater channels to a lateral position on the northern side.

Halsnøy–Huglo area

Fig. 11 shows the interpretation of a sparker profile from the threshold between the islands of Halsnøy and Huglo where the greatest sediment thickness in the entire Hardangerfjord is observed (240 m according to Hoel 1992). Based on air-gun profiles, this area was considered to be a very prominent rocky threshold covered by morainal material (Holtedahl 1975). The acoustic basement is not as clear and continuous as in the Skåneviksfjord due to interference from the first multiple and side-echo effects. The seismostratigraphy is divided into three seismic sequences in the same way as for the Skåneviksfjord. Sequence 1 is ca. 160 m thick and shows even basinfilled reflectors in the western part. The upper part is more wavy, and sloping reflectors may be caused by sediment compaction in the bedrock depressions. An angular unconformity separates Sequence 1 from Sequence 2 with foreset dips increasing from 4° to 12°. Sequence 3 has discontinuous reflectors capping the upper proximal slope. At the base of the distal slope a hummocky surface with tilted blocks (Sequence 4) indicates sliding from the distal slope as well as from the opposite side of the fjord.

Holtedahl (1975) reports a stony proximal slope of the moraine. Distal to the moraine he described max. 240 m thick sediments in the Klosterfjord basin, but the lower part of this is thought to represent Middle Weichselian glaciomarine sediments (Aarseth 1995).

Halsnøy area

Detailed profiling around the island of Halsney shows a thick sediment cover on the proximal as well as on the distal side. Two sets of morainic ridges occur on the proximal side of the island (Fig. 9). A 160 m thick sequence of sediments is found in Høylandsundet distal to the moraine at Sandvoll. In the shallow area further northwest a broad zone with small ridges is found (Hoel 1992), similar to the submarine De Geer moraines in the Sunnmøre area described by Larsen, Longva & Follestad (1991).
Fig. 12. Boomer profile 88–172, southwest of the Halsnøy island showing a sediment sequence disturbed by a glacier coming from the northeast. For location, see Fig. 9.

South of the western part of the island contorted, high-frequency reflectors can be traced to midway between Halsnøy and Hille (Fig. 12). The lower part of the sequence has been subject to compaction while the surface is undulating. Neighbouring profiles have a distinct zone displaying a high amplitude pattern, thought to represent boulders or compaction of glaciomarine sediment by the overriding glacier. Holtedahl (1975), on the basis of air-gun and sparker profiles, described this zone as a continuation of the clayey till on Halsnøy island.

Hardangerfjord area: Summary

The deglaciation in Allerød was extensive in this area and the ice retreated to just north of Ølve (Fig. 1) where seismic profiles show that the radiocarbon dated sub-till sediments described by Aarseth & Mangerud (1974) can also be found in the fjord next to the dated locality (Hoel 1992). The subsequent ice advance was min. 20–25 km and formed the moraine between the islands of Halsnøy and Huglo. From the seismic profiles, there is no indication of an early Younger Dryas ice advance beyond this position as described by Sindre (1980). After deposition of the outermost moraine and distal glaciomarine sediments, the glacier formed some recessional moraines, possibly De Geer moraines, in the shallow areas north and northeast of Halsnøy.

Hardangerfjord–Sognefjord area

Between the Hardangerfjord and the Sognefjord, the submarine Younger Dryas moraines were described on the basis of echosounder profiles (Aarseth & Mangerud 1974), and given the name the Herdla Moraines after the morphostratigraphic type locality on the Herdla Island.

Comments in the following are limited to a few localities with new chronostratigraphic information.

Bjørnefjord–Fusafjord area

A 27 m long core through the sediments at Vinnes, 1 km distal to the moraine deposited by the Younger Dryas glacier in the Fusafjord (Fig. 1) was investigated for sedimentology and bryophytes indicating floods from ice-dammed lakes during the maximum ice-advance in Younger Dryas (Øvstedahl & Aarseth 1975).

Just below the zone of bryophytes, at 8.5–9.5 m depth in the core, shells of Portlandia arctica were found in a zone from 9–12 m depth. AMS datings of the lowermost and uppermost shell of this species gave 10,450 ± 125 BP and 10,345 ± 125 BP (TUa-74 and TUa-75 respectively, Table 3). Fragments of Mya truncata from the basal part of the core gave 11,265 ± 100 BP (TUa-219).

One km distal to the moraine at Strandvik, crushed Balanus balanus in a till-like diamicton were dated to 10,840 ± 190 BP (T-2661). If the field interpretation is correct, the glacier in the Bjørnefjord may have advanced

Table 3. Former unpublished and some published radiocarbon dates mentioned in the text.

<table>
<thead>
<tr>
<th>Locality</th>
<th>14C age</th>
<th>Short description of dated material</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Valen</td>
<td>11,470 ± 180</td>
<td>Shell in till</td>
<td>Holtedahl (1967)</td>
</tr>
<tr>
<td>Ølve</td>
<td>11,230 ± 180</td>
<td>Shell in overconsolidated clay</td>
<td>Aarseth &amp; Mangerud (1974)</td>
</tr>
<tr>
<td>Jektevik</td>
<td>10,600 ± 100</td>
<td>Shell in clay</td>
<td>Genes (1978)</td>
</tr>
<tr>
<td>Vinnes</td>
<td>10,345 ± 125</td>
<td>Shell in core 9.1 m depth</td>
<td>TUa-75</td>
</tr>
<tr>
<td>Vinnes</td>
<td>10,450 ± 125</td>
<td>Shell in core 12.3 m depth</td>
<td>TUa-74</td>
</tr>
<tr>
<td>Austrheim</td>
<td>12,460 ± 100</td>
<td>Lowermost shell outside moraine</td>
<td>T-2107, Gulliksen et al. (1978)</td>
</tr>
<tr>
<td>Austrheim</td>
<td>10,620 ± 170</td>
<td>Lowermost shell outside moraine</td>
<td>T-2106, Gulliksen et al. (1978)</td>
</tr>
<tr>
<td>Fonnes</td>
<td>10,770 ± 140</td>
<td>Shells in glaciomarine clay below till</td>
<td>Aarseth &amp; Mangerud (1974)</td>
</tr>
<tr>
<td>Strangvik</td>
<td>10,540 ± 130</td>
<td>Shells in sand on bedrock</td>
<td>Aarseth &amp; Mangerud (1974)</td>
</tr>
<tr>
<td>Herdla</td>
<td>10,540 ± 130</td>
<td>Shells in sand on bedrock</td>
<td>T-2661</td>
</tr>
<tr>
<td>Mongstad</td>
<td>12,020 ± 120</td>
<td>Shells on bedrock below sediments</td>
<td>T-1606, Gulliksen et al. (1978)</td>
</tr>
<tr>
<td>Mongstad</td>
<td>11,810 ± 120</td>
<td>Shell in clay 1.5 m above bedrock</td>
<td>T-1607, Gulliksen et al. (1978)</td>
</tr>
<tr>
<td>Austrheim</td>
<td>12,460 ± 100</td>
<td>Lowermost shell outside moraine</td>
<td>T-2107, Gulliksen et al. (1978)</td>
</tr>
<tr>
<td>Fonnes</td>
<td>10,770 ± 140</td>
<td>Shells in glaciomarine clay below till</td>
<td>Aarseth &amp; Mangerud (1974)</td>
</tr>
</tbody>
</table>
min. 1 km distal to the moraines at Strandvik without any morphostratigraphic indication. Iceberg ploughing and crushing of the shells cannot be ruled out, however.

**Bjørnefjord**

In the deep Bjørnefjord, where water depths exceeded 550 m in Younger Dryas time, moraine ridges are only found on the north side, outside Strandvik. In a narrow coast-parallel trench here, a lower sequence of high amplitude hummocky reflectors can be traced to a distance of 2 km west of the ridge (Fig. 13). This till may have been formed by the presumed ice advance that deposited the shelly diamicton west of Strandvik and is labelled Sequence 2A. The seismic resolution does not allow a distinction of the structures in the ridge itself, which may represent Sequence 2 and/or 3.

**Fusafjord**

The moraines in the threshold area of the Fusafjord were mapped using a conventional echosounder (Aarseth & Mangerud 1974). Seismic profiles in the same area confirm the interpretation of the position of the moraine ridges. In the rather uneven topography, some of the supposed bedrock knolls turned out to be sediments with a max. 60 m thick structureless sequence sometimes overlying high-frequency acoustically laminated sequences. This is thought to represent till and glaciotectonized sediments overlying glaciomarine sediments, as seen in terrestrial sites in the area (Aarseth & Mangerud 1974).

**Fanafjord**

The relatively low-lying land area along the coast between Os and the Sognefjord partly belongs to the Norwegian Strandflat (Ahlmann 1919; Nansen 1922). The Younger Dryas terrestrial moraines between Os and Herdla are usually small or discontinuous (Aarseth & Mangerud 1974). In the Fanafjord just south of Bergen, an 80 m thick sequence (Sequence 2/3, Fig. 14) in a bedrock depression has several reflectors dipping up-hill in the direction of the ice movement. They are interpreted to be formed during oscillations of the glacier front. The bedrock threshold just outside the moraine might have caused these reflectors, and the units in-between may be considered as a series of stoss-side moraines. The sediments distal to the moraine are divided into three sequences where Sequence 2 is the proximal glaciomarine sediments with zones of point reflectors thought to represent coarse ice-rafted debris (IRD). Sequence 4 is the Holocene sediments which have been penetrated by gravity cores.

**Herdla area**

The island of Herdla is composed mainly of primary and reworked glacial material with local glaciotectonic disturbances suggesting two phases of ice-push in late Younger Dryas (Aber & Aarseth 1988). The highest marine terrace (34 m a.s.l.) consists of 8–10 m of glaciofluvial foreset beds on top of ca. 20 m glaciomarine silt, and a belt of boulders lies across the terrace (Aarseth & Mangerud 1974).

A detailed seismic profile grid was carried out across the Herdleflaket north of the island. High resolution boomer profiles across the moraine (Fig. 15) show two high reflectivity zones with concentrations of boulders on the small ridges at the sea floor. The westernmost ridge has a steep boulder front (13°) from 8 to 14 m water depth, and the high reflectivity seems to continue down to, and below the first multiple. The boulders are interpreted to be deposited at the ice margins during subsequent ice pushes. Between the ridges, apparent horizontal reflectors are caused by instrument ‘ringing’. In this area, loose sand and gravel have been dredged. The sediment
thickness at the moraine 500 m north of the island varies from min. 10 m under the crest of the ridge to max. 100 m on the proximal slope. Distal to the bedrock knoll which is cropping out on the seabed (Fig. 15) is a zone of wavy high-frequency reflectors thought to represent glaciotectonized sediments under, or in front of, the glacier at its maximum readvance. No foreset beds are found in the profiled area.

In the sounds between Herdla and the Fensfjord, 2–3 small, parallel ridges are found in a 1 km wide terminal zone.

**Fensfjord**

The Fensfjord area (Fig. 16), was deglaciated prior to 12,500 BP. At Mongstad, 2 km proximal to the Herdla moraines, a thick sequence of sand is overlain by a clayey till. Barnacles and molluscs on striated bedrock were dated at 12,020 ± 120 BP (T-1606) and molluscs in the clay 1.5 m above bedrock and below 1 m clayey till at 11,810 ± 120 BP (T-1607, Table 3).

The basal part of a sediment section just above sea level at Austrheim, 200 m distal to the outermost moraine was dated at 12,480 ± 100 BP (T-2107) (Mytilus edulis). The uppermost Mya truncata in growing position overlain by 0.8 m silt gave 10,620 ± 170 BP (T-2106) (Gulliksen, Nydal & Skogseth 1978). Shells in glaciomarine clay overlain by till between the moraine ridges at Fonnes were dated at 10,770 ± 140 BP (Aarseth & Mangerud 1974). According to these dates the ice margin in this area reached its outermost position just after 10,600 BP.

The moraine crosses the Fensfjord between a small island west of Mongstad and the island of Sandøy. The 350 m deep sill here consists of max. 240 m thick sediment sequences. Sparker profiles of the moraine have no resolution of the lower parts because of low penetration, even with an air-gun source. The water depths and the relatively narrow fjord channel also make interpretation of this part (Sequence 1) difficult because of side-echoes. Relatively good resolution of the upper parts (parts of Sequence 2 and Sequence 3), however, makes interpretation of the sedimentation dynamics possible (Fig. 17). Depth to acoustic basement is determined from seven transverse profiles. The foreset beds in Sequence 2A have slopes increasing from 2°–6°. Near the crest, between transversal profiles 23 and 24, some reflectors have a gentle up-slope dip. It is not possible to determine whether they represent diamictons or glaciofluvial material. A strong horizontal reflector 80 ms TWT below the surface near profile 26 forms the boundary to Sequence 2B, and a similar reflector 50 ms TWT above forms the boundary to Sequence 3 which consists of horizontal,
high-amplitude reflectors under a small plateau with some relatively steep distal foreset layers (ca. 12° dip).

**Fensfjord moraine: Depositional history**

After the glacier advanced to a position between Mongstad and Sandøy, material was deposited under high pressure building up a moraine of gently dipping foreset layers (Sequence 2A). When the grounding line of the glacier came close to the position of transversal profile 24 the ice advanced and created the reflector between Sequence 2A and 2B. Some high amplitude hummocky reflectors near profile 28 are thought to represent the maximum of this ice advance (1.5 km). After a short retreat to the crest of the moraine with deposition of Sequence 2B (ca. 10°–8° dipping foreset beds), the ice again advanced 1 km and deposited the sediments in Sequence 3. The last ice advance took place after 10,770 BP, and probably also after 10,600 BP.

**Sandøy – Mjømna area**

The area between the Fensfjord and the mouth of the Sognefjord consists of shallow sounds with some large and many small islands. The moraines are relatively large and continuous on the larger islands Sandøy and Mjømna (Undås 1963; Aarseth & Mangerud 1974), but only boulders are found along the shores of the smaller islands. In the sounds, however, up to 80 m thick sequences of parallel-laminated sediments lie distal to one or two ridges of discontinuous reflectors interpreted as sediments deposited directly at the glacier terminus. Figs. 18A and B show the original and interpreted sparker profile west of Mjømna island (Fig. 18). Sequence 1 comprises horizontally laminated reflectors above the acoustic basement of gneissic rocks. The high amplitude zone just below the surface at the base of the steepest slope is thought to represent the coarsest sediment facies deposited near the glacier’s grounding line. Together with its more transparent distal facies, this constitutes Sequence 2. In the ridge itself some weak reflectors display a wavy pattern and dip gently up-slope. They are possibly created by glaciectonic shear movements during the deposition of Sequence 3 near the glacier terminus.

**Sognefjord sill area**

The western 30 km of the Sognefjord has many sills and basins (Holtedahl 1967; Seland 1981; Stensland 1982; Nesje & Whillans 1994). Based on interpretation of the hydrographic chart, Undås (1963) identified one of the outer sills as the most probable Younger Dryas moraine. However, a recessional moraine was suggested by Aarseth & Mangerud (1974) for this 275 m deep threshold. Seland (1981) suggested an Older Dryas age for this threshold with 250 m thick deposits. About 2 km further east, he found a smaller ridge with delta foreset beds. The ridge lies at 400 m water depth and was correlated with the Herdla Moraines. It has been expected that the large Sognefjord glacier would deposit a much larger Younger Dryas terminal moraine. The relatively meagre remains observed may reflect a less stable ice front in the wide and deep threshold area (Seland 1981). Most of the ice and meltwater from the fjord glacier probably drained through the deeper sound branching off northeast of Sula (Fig. 1).

Apart from a 7 km long terminal moraine on Sula island, the position of the Younger Dryas ice margin is
not established in detail between the Sognefjord and the Dalsfjord (Aarseth & Mangerud 1974).

Sunnfjord area

Sunnfjord is the area between the Sognefjord and the Nordfjord, and consists of sounds and islands, small fjords and two longer fjords, Dalsfjord and Førdefjord (Fig. 1). The Younger Dryas moraines bend eastwards in this area (Aarseth & Mangerud 1974), and glacial striae have a general NNW direction revealing an ice flow from the glacier in the Sognefjord to the outer parts of the Dalsfjord (Aarseth 1985).

Dalsfjord

The peninsula at Fure (Fig. 19) has a thick cover of glacial and reworked sediment, and a marine limit of 27 m a.s.l. (Aarseth & Mangerud 1974). The area north of the fjord has extensive clay deposits with abandoned brickyards at Helle and Vårdal. Finds of *Portlandia arctica* have been reported from Helle (Kaldhol 1941). The sediment thickness in the fjord channel is max. 280 m (Risnes 1990).

The Younger Dryas terminal moraine in the fjord is mapped based on 25 transverse sparker profiles (Fig. 19A). A sparker profile crossing the deepest part of the fjord is shown in Fig. 20. Sequence 1 has low continuity. The upper part exhibits divergent acoustic laminae and is interpreted as glaciomarine sediments deposited during the glacial retreat before the Younger Dryas readvance. The lowest 10–20 ms of this sequence has stronger amplitude and may represent coarse glacial material (till or glaciofluvial material).

Sequence 2 displays two separate seismic facies. A chaotic wedge thinning towards the west has parallel, low-amplitude laminae at the base gradually becoming more transparent in the middle part, and an upper part with a hummocky high-amplitude reflection pattern. This is thought to represent a till wedge deposited in front of the grounding line of an unstable glacier. Water depths at the time of deposition were from 450 to 350 m. The irregular surface is probably due to minor oscillations of the glacier front or to gravitational sliding.

The southwestern part of Sequence 2 has hummocky, discontinuous reflectors most likely representing glaciomarine sediments with a high content of IRD. As the glacier front in this area formed a calving embayment, the source of some of these glaciomarine sediments may have been meltwater rivers from channels ending north of Kjøsnes where glaciofluvial foreset beds are present.

Sequence 3 on Fig. 20 has high-frequency laminated reflectors representing distal glaciomarine sediments. The thickness of the Holocene sediments deposited during the glacial rebound (Sequence 4) is probably less than the resolution of the instruments. The largest total thickness of sediments (320 m) is found in the sill area north of the Kjøsnes peninsula (Fig. 19B). The main reason for the shallower fjord here is the top sequence of glaciofluvial foreset beds concentrated in this area where the glacier front almost reached the steep northern fjord side.
The Younger Dryas moraine in the Førdefjord is situated half-way between the mouth and the head of the fjord (Fareth 1970; Kræmer 1977). The 53 m deep sill has a sediment thickness of ca. 110 m with ca. 320 m thick distal sediment units (Risnes 1990). The steep slope of the acoustic basement between the bedrock sill and the basin makes correlations of seismic sequences between the two areas uncertain (Fig. 21).

Sequence 1 is divided into A and B, the lowermost (1A) interpreted to be a diamicton, possibly pre-Late Weichselian in age. Sequence 1B has parallel lamination with highest frequency in the lower part. It most probably comprises distal glaciomarine sediments deposited during glacial retreat in Allerød. 14–16 km proximal to the moraine lies an 80 m thick sediment sequence, showing a strong, gently eastward-dipping reflector in the middle of the sequence. The sediments below this reflector are thought to have been overridden by ice, possibly during the Younger Dryas glacial advance (Risnes 1990).

Sequence 2A of the moraine consists of relatively steep delta foreset beds capped with chaotic high-amplitude reflectors interpreted to be ice proximal diamictons. In the basin this sequence has relatively high frequency and amplitude, and a downlap configuration. Sequence 2B makes up the top of the moraine ridge with parallel foreset beds, and the basin sequence shows an onlap relationship and a wavy reflector pattern thought to represent turbidite channels. Point reflectors may be of IRD origin. The upper sequence (4) fill the channels and forms lobes with tilted reflectors from sliding activity on the south side of the fjord channel.

Sequences 2A and 2B comprise two phases in the formation of the moraine with ice overriding the Sequence 2A foreset beds on the small plateau at 225 m water depth (300 ms TWT). Transverse profiles of the proximal slope show different dipping directions in the two sequences as well as an upper, channelled erosional surface. No sequence is found capping the youngest foreset beds.

**Nordfjord area**

The glacier in the Nordfjord area did split into three fjord glaciers, all three depositing large moraines (the Nor moraines) during the middle to late Younger Dryas (Fareth 1970, 1987; Mangerud et al. 1979; Rye et al. 1987). After retreating in Allerød the northernmost glacier advanced min. 35 km in early Younger Dryas before it retreated 15 km and deposited the sandur-delta at the Nor type locality (Klakegg & Nordahl-Olsen 1985). No terminal moraines from the supposed outer-
most early Younger Dryas terminus are found in the fjord (Giskeødegaard 1983). The southernmost terminal moraine is situated near the head of Gloppen fjord, while the moraine in the main fjord (Hundvikfjord) forms the sill between Lote and Anda (Giskeødegaard 1983; Figs. 22, 23). The marine limit in this area is 50 m a.s.l. (Fareth 1987).

Lote–Anda area (Hundvikfjord)

A sharp ridge, 100–130 m high, with a distal slope of max. 17° forms the sill in the 2 km wide fjord (Fig. 23A). The sediment thickness is largest on the south side of the proximal slope leaving a channel to the north (Fig. 22). Fig. 23A shows a sparker profile along the direction of deposition with interpretation in Fig. 23B. The proximal part of Sequence 1 is semi-transparent, while the distal part, west of the bedrock sill, has a high amplitude, parallel lamination in the lower part gradually decreasing upwards. The sequence is interpreted as glaciomarine sediments deposited during glacial retreat in Allerød. Sequence 2A has high amplitude reflectors and a very hummocky western part, while the regular foreset beds make up Sequence 2B with slopes increasing from 5° to 15°. The contorted area below the crest (Sequence 3) is interpreted to be a max. 50 m thick diamicton. Similar reflectors in the lower part of the proximal slope most likely belong to the same sequence. Sequence 4, just below the uneven surface at the base of the distal foreset beds, represents slide material. Bottom photographs from the top of the moraine show stones on sandy gravel, while sand and gravel lie on the proximal and distal slopes (Giskeødegaard 1983).

After a retreat of unknown extent during Allerød the glacier in the main fjord advanced to deeper water at the intersection with Gloppen. A 2.5 km retreat followed to a more stable position where foreset beds started to build up. A subsequent ice-push deposited the diamicton capping the morainal crest. Transverse profiles show the asymmetry of the proximal slope of the deposit with a >100 m deep erosional channel on the northern side.

The bottomset beds of Sequence 2 are thin just west of the moraine because of the profile orientation (Fig. 22A), but increase to 80 m 5 km further west where the total sediment thickness amounts to 400 m (Giskeødegaard 1983).

Gloppen

The sediment sequences in Gloppen (Fig. 24) can be interpreted in a similar way to those in the Hundvikfjord. The glacier retreated in Allerød time and deposited glaciomarine sediments above bedrock (Sequence 1). The first Younger Dryas ice advance can be traced down to 350 ms TWT where Sequence 2A was deposited in ca. 300 m water depth. This was followed by a 2 km glacial retreat to a stable position from where the foreset beds started to build up (Sequence 2B).
A subsequent ice advance deposited a diamicton (chaotic reflection pattern, Sequence 3A) on top of the foreset beds and caused a sliding of glacial material to 1.5 km from the base of the foreset beds. After a short retreat, the last ice advance deposited the upper diamicton (Sequence 3B). Sequence 4 is younger glaciomarine sediments and Holocene prodelta sediments with a possible turbidite channel also found on the transverse profiles.

Sunnmøre area
The area between the Nordfjord and the town of Ålesund (Fig. 1), known as the Sunnmøre area, is characterized by an alpine topography, and during Younger Dryas local glaciers occupied numerous cirques (Reite 1967). A regional map of the main features of glacial geology of Central Norway shows the approximate position of the main ice margin in Younger Dryas (Sollid & Sorbel 1979). A Younger Dryas valley glacier from the Nordfjord area drained over a mountain pass and down to the Sunnylvsfjord through Sunnylvsbygdå valley (Fig. 25; Fareth 1987; Rye et al. 1987), while the glacier terminus in the Geirangerfjord was located near the mouth of the fjord.

In the steep mountain sides along the innermost fjords and valleys almost no lateral moraines from the inland ice are preserved. The best way of mapping the extent of the Younger Dryas glacial advances in this area is by reflection seismic profiling (Giskeødegaard 1983). Later, more detailed profiling has revealed glacial oscillations during Younger Dryas. The Sunnmøre area has been subject to rock sliding during Late Weichselian and Holocene (Jørstad 1968; Blikra & Nemec 1993; Blikra 1994).

Sunnylvsfjord and Geirangerfjord
An interpretation of the depositional history for this area is based solely on detailed seismic profiling (Figs. 26, 27). During deglaciation of the fjord, a large rockfall from a 1500 m high mountain was deposited across the fjord east of Åkernes (Fig. 25), where hummocky and high amplitude reflectors are found below ca. 25 m of Late Weichselian and Holocene sediments.

After deglaciation of the fjord system the glacier advanced to a position east of Ljøvika (Fig. 25). This advance produced a till unit that shows up as a high amplitude reflector above bedrock (Figs. 26, 27). Material from a giant rockfall on the east side of the glacier was most likely transported a short distance towards the glacier terminus and deposited on the fjord bottom where it created hummocky high amplitude reflectors (Fig. 26).

The unstable glacier terminus with extensive calving in the Sunnylvsfjord led to a retreat of the glaciers in both fjord branches. The Sunnylvsfjord glacier retreated to the bedrock sill at Bjørkeneset where a terminal moraine with foreset beds was formed. Thick glaciomarine sediments were deposited on top of, and beyond, the till and rockfall material (Fig. 26).

The glacier in the Geirangerfjord retreated to a position east of Stabbeflura, and more than 80 m thick
glaciomarine sediments were deposited in front of the glacier before it gradually advanced. Glacial and glaciofluvial erosion took place at the proximal slope while meltwater brought coarse material to the grounding line which prograded to the south of Lundanes. The fine material was transported as sediment gravity flows down the turbidite channels and partly covered the deposits from the Sunnylvsfjord (Fig. 26). After deglaciation of the area, rockfall activity from the mountain Nokkenibba (1380 m a.s.l.) covered parts of the moraine in Geirangerfjord (Fig. 27), while relatively coarse sediments filled most of the basin proximal to the inmost Sunnylvsfjord moraine.

Norddalsfjord

The terminal moraine in Norddalsfjord (Fig. 1) does not form a conspicuous ridge on the fjord bottom because it is buried below younger glacial and prodelta sediments from the Valldal valley (Fig. 28). The sediment sequence is interpreted as follows: The lower part of the distal sediments was deposited during the first part of the deglaciation with a thin unit of coarse sediments capping the bedrock in the basin. When the glacier retreated beyond the eastern rock threshold, deposition decreased and the sediments underwent compaction that created a relatively strong seismic reflector. An unstratified coarse sequence was deposited at the threshold before the foreset layers started building a prominent ridge across the 2 km wide fjord.

On the south side of the fjord, parts of the ridge have been subject to sliding, creating slide material distal to the delta front. Slides have also taken place further west as indicated by the western slide scar, leaving a sediment ledge along the fjord side. There is no acoustic indication of an ice advance following the deposition of the foreset beds in the Norddalsfjord. Distal sediments from ice-contact deposits near the mouth of the Valldal valley fill the basin created by the moraine. Holocene prodelta sediments from the Valldal river rest on top of these sediments.

Romsdalsfjord

The Younger Dryas moraine in the Romsdalsfjord is partly buried by the recent river delta at Åndalsnes (Sollid & Sorbel 1979). A profile along the fjord 1 km north of the delta revealed a submarine ice-front deposit with glaciofluvial foreset beds but, due to the freshwater input from the river, the sparker records were too poor to make a detailed interpretation of this deposit (Aarseth, Lønne & Giskeødegaard 1989).

Stratigraphy of the Sequences (1 to 4)

Reconstruction of the sedimentary environments under which the sequences are deposited is shown in two series of models (I and II) representing deposition in deep (> 100 m) and shallow water respectively (Fig. 29). The succession through the different stages of the models has been labelled according to the seismic sequences so that 1 to 4 represent Sequences 1 to 4.

Sequence 1

Sediments older than Late Weichselian cannot be identified from the seismic stratigraphy alone. High-amplitude reflectors just above the acoustic basement are, therefore, with one exception, interpreted to represent coarse glacial sediments deposited as meltout of basal and englacial debris during the deglaciation in Allerød, and incorporated in Sequence 1. In Førdefjord (Fig. 21) the lowermost sequence of the distal basin may represent a pre-Late Weichselian diamicton and is labelled Sequence 1A.

Sequence 1 is generally acoustically laminated and contains glaciomarine sediments deposited in front of the retreating glaciers. The bulk of these sediments is thought to settle from flocculation within a few km from the glacier margin where the sediment-loaded meltwater meets the brackish fjord water. This may take place as underflows (Powell 1981), although the existence of underflows is strongly questioned by Syvitski (1989) who reports deposition of 70% of the sediment load by overflow within the first 500 m of an ice front. The sedimentation rates in modern glacier-fed fjords decrease exponentially with distance from the source (Relling & Nordseth 1979).

The acoustic lamination may represent mud interbedded with silt and fine sand laminae settling from rhythm-
The seismic configurations of Sequence 1 vary in thickness from less than the instrument resolution at the lateral trench outside Strandvik (Fig. 13), to 300 m 5 km distal to the moraine in Nordfjord. Sequence 1 is also found proximal to the moraines in some areas, and may indicate a minimum retreat of the glaciers in Allerød before the Younger Dryas ice advance.

In Jøsenfjord, Sequence 1 is very thick just below the moraine ridge (Figs. 5, 6). A model for this sedimentation was presented by Austbe (1988) who postulated an ice front at the head of the fjord, and a heavy flocculation of glaciomarine sediments at the mouth, where the transition between the fjord water and the coastal water caused special hydrographic conditions that generated flocculation.

### Sequence 2

The seismic configurations of Sequence 2 are more variable than Sequence 1. This is expected because of the great variety of processes taking place at the grounding line producing many different lithofacies (Powell 1981, 1984). Terrestrial evidence also shows that the Younger Dryas glacier front in western Norway was unstable, with several ice-push phases (Bløstad & Anundsen 1983; Rye et al. 1987; Aber & Aarseth 1988).

The glacier remobilized some of the Allerød sediments during its readvance in early Younger Dryas. In some areas, such as the Dalsfjord (Fig. 20) and the Nordfjord area (Figs. 23, 24), Sequence 2 sediments interpreted to represent an ice-push are found overlying Sequence 1. In most of the investigated deep fjords, Sequence 2 represents delta foreset beds from glaciofluvial sedimentation (Fig. 29). The lowermost layers are less distinct, but have angles varying from 2°-6°. These low-angle layers may represent high-energy meltwater regimes or material being squeezed out at the grounding line. The high clay content of the overridden deposits of Sequence 1 would facilitate such squeezing to produce a diamicton as envisioned for these lowermost foreset beds. Similar low-angle foreset beds are observed in the seismostratigraphy of Younger Dryas moraines in North Norway (Andersen et al. 1982) and in terrestrial sections (Andersen et al. 1981; Lønne 1993). Sections in proximal slopes of Younger Dryas moraines in the investigated area also show diamictons (Aarseth & Mangerud 1974; Holtedahl 1975).

Moraine-proximal up-slope dipping reflectors in Sequence 2 are found in a few locations such as the Fensfjord (Fig. 17). These are thought to be glaciofluvial sediments deposited in tunnels between the glacier terminus and the moraine. At some locations erosion has taken place, creating channels and angular unconformities. Similar structures are found in Younger Dryas moraines in eastern Norway (Brandal & Heder 1991).

The foreset beds of the terminal moraines become gradually steeper, the steepest layers dipping from 9° to 17°. In some fjords a major unconformity is seen between two main phases of glaciofluvial deposition. In the Fensfjord and the Førdefjord (Figs. 17, 21), the glaciers advanced over Sequence 2A before they retreated and deposited Sequence 2B. In both cases the foreset beds become steeper during the younger sequence. In shallow fjords and sounds the foreset beds are thin or missing. At Herdla foreset beds are found on the island, but not on seismic profiles across the submarine moraine just north of the island, because the island lies at the mouth of the Herdla fjord which guided the meltwater flow.

### Sequence 3

In most of the investigated fjords a sequence of chaotic reflectors caps the foreset beds of Sequence 2 (Fig. 29). This represents ice-pushed or meltout material in addition to supraglacial material dumped from the ice terminus (Powell 1981). North of the Herdla island, for example, (Fig. 15) this sequence comprises two or more distinct ridges that contain stones and boulders on and below the surface. In Gloppen, Sequence 3 is subdivided into A and B, representing two separate ice-push phases (Fig. 24).
Up-slope dipping reflectors are also seen in Sequence 3 in some of the moraines. These are thought to be formed by shear movements near the glacial terminus (Fig. 18), or from meltout till being plastered on the stoss side (Fig. 14). On some moraines the proximal slopes have reflectors lying parallel to the surface. The instrument resolution makes it difficult to determine whether this represents glaciofluvial material, till, or younger, redeposited material (Seq. 2, 3 or 4).

Sequence 4

Sequence 4 incorporates all sediments deposited after the last observed ice-push in Younger Dryas and comprises slide material, thick proximal basin infill or reworked sediments. In many of the fjords, however, slides on the delta fronts may have been caused by ice-push over unstable foreset beds. In some cases this has created delta front slopes steeper than the youngest foreset beds.

Moraines deposited in shallow water in sounds exposed to tidal currents or wave action commonly show evidence of lag formation and redeposition (Figs. 15, 29). Estuarine circulation over fjord sills (Svendsen 1977) may cause erosion on top of the moraines, even in deeper water, causing a transport of sand mainly down the proximal slope.

Sediment volumes

In some of the fjords, sediment volumes are calculated for each of the various sequences and facies (Table 2). Volumes are estimates for the general area confined to the moraine and should not be considered representative of the entire fjord. Some deductions, however, can be drawn from these data. The coarse part of Sequence 2 (mainly glaciofluvial material) represents a relatively small portion of the total sequence (usually <10%). In comparison, the portion of bedload in recent glacial rivers in western Norway is 30–50% (Østrem, 1975). One reason for this difference may be that most of the Sequence 1 material (clay and silt) proximal to the Younger Dryas moraines was eroded and redeposited. The crushing and abrasion by the large Late Weichselian glaciers may also have differed considerably from that by the smaller recent glaciers.

Glacial dynamics as inferred from the seismic stratigraphy

The glacial dynamics during the formation of the Younger Dryas moraines are visualized in the models presented in Fig. 29. South of Cape Stad there is evidence for glacial retreat to the heads of Gloppen (Fig. 24) and lake Hornindalsvatn, east of Nor (Klakegg & Nordahl-Olsen 1985).

Whether the glaciers in the larger fjords further south retreated to the heads of the fjords during Allerød is questionable. According to the model of Alaskan tidewater glaciers (Meier & Post 1987), glacial calving in an inlet must proceed until the terminus reaches shallow water at a sill or at the head of the inlet. From there the glacier can start advancing by building a morainal bank, thereby reducing calving, and the glacier may advance and transport the morainic material forward. Andrews (1990) suggests that substantial calving filled the fjords in Arctic Canada with icebergs, and seasonal fast-ice together with the jamming of the icebergs inside a sill reduced the calving, making a new glacial advance possible without building a morainal bank. This model seems to fit the deep western Norwegian fjords. In contrast, the application of the Alaskan model for the Hardangerfjord would imply that morainal banks were built and transported min. 100 km forward in a short time in more than 1200 m water depth. Even with the known min. 25 km retreat in Allerød, morainal banks had to be formed at 700 m water depth and transported forward. The Younger Dryas moraine between Halsnøy and Huglo is too small to fit this model.

The area south of Stad (Fig. 1) had glaciers on mountain plateaus close to the coast. A depression of the equilibrium line altitude (ELA) during early Younger Dryas produced large areas above this line (Mangerud et al. 1979) that led to advancing fjord glaciers in this area, while local glaciation was more extensive further north (Reite 1967).

The Younger Dryas glaciers advanced to positions where they either maintained equilibrium between ice flow and calving, which was the most important ablation factor, or after quick advances retreated to positions where such equilibrium was obtained (Fig. 29). The fjord depths as well as the fjord widths control the rate of calving (Brown, Meier & Post 1982). This can be seen from the locations of many of the moraines on bedrock sills or close to fjord intersections with abrupt changes in widths and water depths.

At these quasi-stable ice fronts glaciofluvial ice contact deltas were deposited and this led to a decrease in water depth that then reduced the rate of calving. This could enable the glaciers to advance and deposit diamictons on top of the sorted foreset beds. Another possible reason for these glacial advances could be a continued positive glacial mass balance. This seems to be the case in areas where calving played a minor role (shallow sounds, fjord heads or terrestrial sites) where two or more ice-push phases are observed (Fig. 29). A Younger Dryas glacier advance in the order of 10 km, and most likely 17–18 km, has recently been reported from the Kragere area (Fig. 1) south-eastern Norway (Bergstrøm 1995). This is thought to be caused by an active ice dome on the southwest part of the Hardangervidda mountain plateau. Asymmetric sedimentation and/or erosion have been shown in fjords where the moraines were deposited close to a widening of the fjord, or where tributary glaciers
coalesced near the terminus. Erosion and/or the sedimentation of channels on the proximal slope are found in similar settings and may be due to diffuseness with the subsequent formation of longitudinal crevasses at the glacier's terminus.

The problem of dating the Younger Dryas moraines in western Norway

Mangerud (1980) states that the moraines were probably deposited closer to 10,500 BP than to 10,000 BP. In a recent paper, Mangerud favours a maximum glacial readvance in late Younger Dryas (Andersen et al. 1995). This is based on two early 14C dates with high standard deviations from Os (Holledahl 1964). The new dates presented here from Vinesse in the Fusa fjord, just across the fjord from Os, support Mangerud's (1980) suggestion (Table 3). The problem of 14C dates in this chronzone with a plateau in the correlation of 14C to calendar years, however, makes a precise dating difficult (Bard et al. 1993).

The Vedde ash, now dated to 10,300 BP (Bard et al. 1994) could be a useful tool, but so far it has not been found in sections in connection with the moraines of the outlet glaciers from the main ice sheet. The retreat from the Younger Dryas moraines was extremely rapid as the innermost fjords were ice-free some hundred years later (Andersen 1980).

Conclusions

The Younger Dryas ice margin in western Norway crossed numerous fjords and sounds. The seismostratigraphy of the corresponding moraines show that they vary in size and internal structure. Similarities in structure can, however, be used to present two models of sedimentation and glacial dynamics at the terminus of the Younger Dryas fjord glaciers (Fig. 29). In deep fjords the glacier advanced to an unstable position before it retreated 1–3 km to build up an ice-marginal delta. A positive mass balance of the glacier, or reduced water depth as a result of deposition at the margin of the glacier, led to glacial advance and deposition of a diamicton on top of the forested beds. In shallow water the glacial oscillations were most likely controlled by changes in the supply of ice from the mountains, and two or three moraine ridges may be found, often with large boulders along the supposed ice margins. The coarse-grained fraction of the sediments deposited by the mapped fjord glaciers during Younger Dryas represents only ca. 10% of the total sediment volume. This leads to the conclusion that the material transported by the meltwater during this phase was mainly the product of the glacial scouring and reworking of glaciomarine Allerød deposits.

References


References


The Younger Dryas was extremely rapid as the innermost fjords were ice-free some hundred years later (Andersen 1980).

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