FIELD EXCURSION GUIDE

Changing Glaciers:

SYMPOSION
FJÆRLAND
NORWAY
JUNE 1996
Changing Glaciers:

Field guide to excursions in conjunction with symposium in Fjærland Norway, June 1996

by
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Edited by N. Haakensen
Preface

This field guide is made for the excursions arranged in conjunction to the Symposium on Changing Glaciers arranged in Fjærland, Norway June 24-27, 1996. The symposium is sponsored by the International Glaciological Society (IGS), Cambridge, England, the Norwegian Glacier Museum, Fjærland, the Norwegian Polar Institute (NP) and the Norwegian Water Resources and Energy Administration (NVE)

Responsible for organizing the excursions are Nils Haakensen, O.M.Korsen, Atle Nesje and Rune Aa. Many other persons, however, are involved in the arrangement.

There are four separate excursions

  Full day excursion to Brigsdalsbreen and Loen on June 23.  
  Responsible Atle Nesje.

  During the symposium there will be a half-day tour to the glaciers in Fjærland.  
  Responsible O.M. Korsen.

  Full day excursion to Veitastrond and Austerdalsbreen on June 28.  
  Responsible Rune Aa.

  Full day excursion to Jostedal and Nigardsbreen on June 29.  
  Responsible Nils Haakensen.

The individual chapters in the guide are listed on the Page of Contents.

The guide has been edited by Nils Haakensen, who also did the technical editing.

Following publications are enclosed with this guide.

1)  Norges vassdrags- og energiverk: Map of Nigardsbreen. Scale 1:20 000.

2)  Aa, A. R. 1988: Brigsdalsbreen, 1318 II  
    Kvartærgeologisk kart (Quaternary map), Scale 1:50 000. Norges geol. unders.
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The travelling routes are shown by red lines

Scale, 1 : 350 000

(Geographical data: Statens Kartverk N250)
Glaciation and deglaciation of Nordfjord, western Norway
Atle Nesje, Department of Geography, University of Bergen

Topography and bedrock
In the Nordfjord area (Fig. 1), a high mountain plateau is incised by fjords and deep valleys. The summit areas are relatively flat and undulating, commonly bordered by steep cliffs. The high mountain plateau around Nordfjord descends from about 1800 m elevation at Jostedalsbreen to 400-500 m at the coast. The bedrock is dominated by gneiss with some massive granite, amphibolite, quartzite, schist, eclogite and dunite. The bedrock in the Ålfotbreen area south of Nordfjord consists of Devonian sandstones and conglomerate.

The last glacial maximum
The highest mountain areas around inner Nordfjord are covered by in situ blockfields and several mountains exhibit an alpine, rugged morphology. The term blockfield is used for autochthonous, angular boulders formed through weathering of the local bedrock. The blockfields commonly have sharp lower boundaries. The weathering boundary is defined as the level between the lowest summits with blockfields and the highest summits showing evidence of glacier overriding (striations, erratics etc.). In inner Nordfjord, the weathering limit descends north-westward from Jostedalsbreen (ca. 1750 m) with an average gradient of about 7 m/km to approximately 1500 m in the mountain areas between inner Nordfjord and Sunnmøre (Fig. 2). Rock surfaces between the blockfield boundary and the Younger Dryas (11,000-10,000 radiocarbon yrs BP) lateral moraines show signs of glacial erosion, without any blockfield formation, despite that they have been exposed to subaerial weathering for at least 11,000 radiocarbon years. The weathering boundary is up to 600 m higher than the Younger Dryas lateral moraines (Fig. 2). Nesje et al. (1987) and Rye et al. (1987) concluded that the weathering boundary in inner Nordfjord represents the upper level of inland ice during the last glacial maximum. It is, however, possible that the gently undulating mountain
Fig. 1. Location map, Nordfjord. Adapted from Dahl and Nesje (1992).

Fig. 2. The blockfield boundary in inner Nordfjord. From Nesje et al. (1987).
Fig. 3. Time/distance diagram for the sequence of deglaciation of Nordfjord. Adapted from Sønstegaard et al. (in prep).
plateaux above the weathering boundary were covered by thin, cold-based and dynamically inactive local snow fields or minor ice caps unable to erode the blockfields already formed.

Deglaciation

In the coastal area, a series of terminal ridges from the deglaciation period has been mapped on Stadlandet (Fig. 1), of which the Ferstad moraine is the most prominent (Mangerud et al., 1979). Radiocarbon dating of limnic sediments from a lake at Kråkenes showed that the coastal area of Nordfjord was deglaciated before 12,320 ± 120 radiocarbon years BP (Mangerud et al., 1979) (Fig. 3). At the southern side of Nordfjorden, east of Davik, Fareth (1987) mapped lateral moraines, which he interpreted as marking a former ice-front position somewhere in the outer fjord area, termed the Davik Stadial, of probable late Bølling/Older Dryas/Allerød age. Recent investigations may, however, indicate that these moraines were formed by the Younger Dryas Ålfotbreen plateau glacier (Sønstegaard et al., in prep.) Fareth (1987) also mapped lateral moraines in the mountain areas in middle and inner Nordfjord, named the Vardehaug moraines (Fareth, 1987).

The continental Scandinavian ice sheet retreated several tens of kilometres before the Younger Dryas. During the Younger Dryas, however, the retreat halted and along a considerable part (mainly in western Scandinavia) a readvance of tens of kilometres took place. West Norwegian fjord glaciers advanced about 50 km (e.g., Andersen et al., 1995). South of Sognefjorden, the Younger Dryas ice sheet nearly reached the outer coast with no local ice caps beyond the ice sheet. North of Sognefjorden, however, the ice margin of the ice sheet continued inland. In Nordfjord the ice margin crossed the middle part of the fjord, and in the Møre area to the north, the ice margin was located at the head of the main fjords (Fig. 4). North of Sognefjorden numerous cirque glaciers and an ice cap (over the Ålfoten area) existed beyond the outlet fjord and valley glaciers from the main continental ice sheet (Fareth, 1987; Dahl and Nesje, 1992; Sønstegaard et al., in prep.). Fareth (1987) mapped some prominent terminal and lateral moraines (The Nor moraines) of Younger Dryas age deposited
Fig. 4. The location of the Younger Dryas ice margin in southern Norway. Adapted from Andersen et al. (1995).

Fig. 5. Marine levels in Nordfjord. Modified from Fareth (1987) and Rye et al. (1987).
by the valley/fjord glaciers in middle and inner Nordfjord. Mangerud et al. (1979) dated the Younger Dryas Nor readvance to have culminated close to 10,750 radiocarbon years BP.

The retreat of the valley and fjord glaciers from the position of the Nor moraines to the valley mouths in inner Nordfjord was, according to radiocarbon dates and marine levels (see later), relatively rapid, primarily because of effective calving of the fjord glaciers (Rye et al., 1987). In each of the valley-mouths of the Stryn, Loen, and Olden valleys are two ice-marginal deposits (termed the Vinsrygg and Eide moraines) a few kilometres apart, indicating minor readvances or halts in the general retreat. The ice-frontal deposits are located on bedrock thresholds and narrow parts of the fjord/valleys. The formation of the ice-marginal deposits may, therefore, be explained in terms of glacial dynamics as a result of rapid calving of the fjord glaciers and unstable, dynamically active valley glaciers. During the subsequent stabilising process, when the glacier fronts became grounded, the glaciers either readvanced or at least halted their general retreat. The final deglaciation was characterised by retreating and downwasting glaciers.

Marine levels

The Younger Dryas shoreline (isobase direction 30° NE) is entirely submarine in the coastal area (Fig. 5). Fareth (1987) showed, with a few exceptions, one marine level between the Tapes level in outer Nordfjord and the Nor moraines, with an average gradient of 1.06 m/km. In the valley mouth of Olden there are marine deposits just below the extrapolated Younger Dryas sea level, which indicate either a rapid calving of the fjord glacier to inner Nordfjord in the late Younger Dryas or during the early Preboreal, or deglaciation in a period of relatively stable sea level before, or in the early period of a rapid regression recorded in the Preboreal Chronozone. Gradually falling marine levels are recorded in the Olden, Loen, and Stryn valleys. Some striking differences between the three valleys are, however, recognised. The largest decline in the marine levels is recorded in the Stryn valley, while less declines are recorded in the Olden valley (Fig. 5). This indicates an earlier deglaciation in the Olden valley.
than in the Stryn valley, also supported by the general knowledge of the sequence of
deglaciation, with downwasting, dynamically inactive glaciers in the Lovatnet and
Strynevatnet basins in the latest phase.

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Younger Dryas glaciation in the Ålfoettbreen area, western Norway; Evidence from lake
sediments and marginal moraines.
Holocene lake-level variations of Jølstravatnet

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This chapter is based on Klakegg and Rye (1990).

The lake Jølstravatnet and its surroundings demonstrate the effect of glacio-isostatic rebound on lake-level development. Jølstravatnet (40 km²) is 29 km long and has a maximum water depth of 233 m. Its present outlet lies at Vassenden in SW. The outlet river (Jølstra) passes a rock threshold which has a veneer of boulders. The level of the threshold is at 205.9 m a.s.l. The threshold at the NE end of the lake (at Skei) is only ca. 3 m higher than the present outlet threshold.

Jølstravatnet is located within the Younger Dryas (Nor moraines) ice margin (Fig. 1). The lake area was deglaciated during the Preboreal Chronozone (10,000-9000 radiocarbonyrs BP). The isobases cross the lake at almost an oblique angle (Fig. 2).

The lake history can be subdivided into five phases (Fig. 3). Approximate radiocarbon ages for the drainage phases are obtained by the construction of a time-gradient curve for the area.

During phase I (prior to 9500 BP) the eastern part of Jølstravatnet was occupied by glaciers.

Phase II (9500-9200 BP) started with an outburst from the lake and a 12-m lowering of the lake level. From the new lake level, the eastern outlet at Skei was established.

During phase III (9200-7500 BP) glacio-isostatic rebound caused a fresh water transgression in the western part of the lake.

Phase IV (7500-6000 BP) was characterised by outlets at both ends of the lake. The drainage through the NE outlet stopped about 6000 BP, marking the transition to phase V.

During phase V the water flowed (as at present) through the SW outlet at Vassenden.

References


Fig. 1. Location map. Adapted from Klakegg and Rye (1990).

Fig. 2. Terrace levels (m) along Jølstravatnet. At Vassenden and Skei the elevations of the thresholds are given. Isobases from Fareth (1987). Adapted from Klakegg and Rye (1990).
Fig. 3. Lake drainage phases I-V. Modified from Klakegg and Rye (1990).
Holocene glacier fluctuations in the Jostedalsbreen region
Atle Nesje, Department of Geography, University of Bergen.

Early Holocene
The sequence of deglaciation during in the Jostedalsbreen region (Fig. 1) during the mid Preboreal Chronozone (9500 ± 200 radiocarbon years BP) was characterized by vertical wastage, indicating that the ELA was at, or above the summit plateaux (Rye et al., 1987; Nesje, 1992) (Fig. 2). The valleys surrounding the Jostedalsbreen ice cap were deglaciated during the latter half of the Preboreal Chronozone (9500-9000 yr BP) (Rye et al., 1987; Nesje et al., 1991; Nesje, 1992). In the late Preboreal/early Boreal (9100 ± 200 yr BP), however, marginal moraines were formed up to 1 km beyond the 'Little Ice Age' moraines deposited in front of the present valley outlet glaciers discharging from the Jostedalsbreen ice cap (Fig. 3). Inferred from the upper altitude of lateral moraines formed during this readvance, and calculations of the ELA depression based on an accumulation-area ratio (AAR) of 0.6, the average ELA lowering is calculated to 325 m (Nesje, 1992).

The Holocene thermal optimum
After 9000 yr BP a rapid increase of pine (Pinus sp.) occurred based on palynological studies (Kvamme, 1984) at Sygneskardet, Sunndalen (Fig. 1). The pine maximum occurred close to 8400 yr BP, suggesting that modern temperatures were achieved close to the Preboreal/Boreal transition (9000 yr BP). The pollen record also shows that between about 6500 and 5000 yr BP stands of elm (Ulmus sp.) grew close to the site, which at present lies at the birch (Betula sp.) forest limit. Forest stands of elm and birch need minimum summer (June-September) temperatures of about 11 and 7 °C, respectively (Dahl, 1967), indicating summer temperatures close to 4 °C warmer than at present close to the present glacier during the Atlantic Chronozone. Without the modern cooling effect perceptible at the birch forest limit at Sygneskardet, however, the difference between the Atlantic (8000-5000 yr BP) and the modern mean summer temperature is calculated to 1.5-2.0 (1.8) °C (corrected for glacio-isostatic recovery) (Nesje and Kvamme, 1991; Nesje, 1992).
Fig. 1. Map showing the Jostedalsbreen region.
Fig. 2. Holocene equilibrium-line altitude (ELA) fluctuations in the Jostedalsbreen region. Adapted from Nesje and Kvamme (1991).

Fig. 3. Holocene glacier fluctuations in the Jostedalsbreen region. Pattern indicates inferred extent of ice cover. Horizontal scale is schematic. Adapted from Nesje and Kvamme (1991).
Radiocarbon dates from the base and top of a para-autochthonous peat exposed by recent (AD 1962-1966) retreat of Tunsbergdalsbreen (Fig. 1) demonstrate that the glacier front terminated continuously upvalley between 8083 ± 100 and 3855 ± 55 yr BP (Mottershead et al., 1974; Mottershead and Collin, 1976). This may suggest that the ELA was as high as, or higher than at present during 45% of the last 9000 years.

'The Little Ice Age'

The first historically reported damages by the 'Little Ice Age' glaciers occurred in AD 1339 (Grove, 1988). Several farms around the ice cap suffered severely from glacier advances and associated avalanches, rockfalls, and landslides in the 17th and 18th centuries (Grove and Battagel, 1983). The earliest definite evidence of damage to farmland by advancing glaciers comes from a brief account dated AD 1684. The 'Little Ice Age' glacier advances culminated during the mid 18th century [Nigardsbreen (Fig. 1) in 1748 AD]. A regional lichenometric dating study of seven outlet glaciers from the Jostedalsbreen ice cap, suggests that four glaciers reached their maximum 'Little Ice Age' maxima prior to AD 1780 (Bickerton and Matthews, 1993). Inferred from the upper limit of lateral moraines combined with an AAR of 0.6, reconstructed glaciers during the maximum 'Little Ice Age' advance in the mid 18th century suggest an ELA depression calculated to 100-150 m (Torsnes et al., 1993), indicating a mean summer-temperature lowering of 0.6-1.0 °C (Fig. 3).
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The Late Weichselian ice sheet in the Nordfjord-Sunnmøre area and deglaciation chronology for Nordfjord,

Modern and Little Ice Age equilibrium-line altitudes on outlet valley glaciers from Jostedalsbreen, western
Norway: an evaluation of different approaches to their calculation. Arctic and Alpine Research 25, 106-116.
The glacier destruction of the Tungøyane farm in Oldedalen, the 12th December 1743

Atle Nesje, Department of Geography, University of Bergen.

During the 'Little Ice Age', advancing glaciers around Jostedalsbreen caused severe damage to pastures and farmland (Rekstad, 1902; Eide, 1955; Grove and Battagel, 1983; Grove, 1988). As a result of severe climatic conditions during this period, several farms in inner Nordfjord were also exposed to other types of natural hazards (floods, rock falls, snow avalanches etc.).

Historical evidence shows that the advances of Nigardsbreen in Jostedalen and Brenndalsbreen in Oldedalen led to the most severe damage, and that which affected Tungøyane was the most tragic. The destruction of Tungøyane took place over a period of about 40 years, when the glacier front of Brenndalsbreen was situated in the mouth of Brenndalen (Fig. 1), causing a series of avalanches and floods over the farmland.

Damage to Tungøyane

From 1702 and onwards the Tungøyane farm was regularly damaged by floods and snow avalanches, and the farmers and their families had to move out of their houses during the worst avalanche periods. In 1723 it was stated that the farm was easy to run, but it was situated in front of an advancing glacier (Brenndalsbreen).

During a tax inspection on the 12th of October 1728, the court stated that the two farmers, before the tax reduction in 1702, paid their taxes according to the instructions from the King and the Church. However, in the late 1720s, the farmers were not able to pay their taxes due to severe damage. Brenndalen, the valley above the farm occupied by the advancing glacier, had previously been good pasture land for cattle. In addition, the farmland around the houses was regularly covered by boulders, sand and gravel from river floods. In 1728 they therefore had to move the houses away from the river plain to a place where they felt safe (2 in Fig. 1).
Fig. 1. The Tungeyane - Brenndalsbreen area in Oldedalen, Nordfjord.
In the middle of summer 1733 the farmland once again suffered severe damage by floods from the glacier. On the 2nd November 1734 the court (7 persons), led by U. Kås, visited the farm to estimate the damage. At that time, the glacier tongue had advanced through a narrow canyon just above the buildings. The glacier, "that never will disappear", they stated, had advanced down into the main valley. At that time the main river in Oldedalen also changed its course, running over what previously had been their best farmland. In 1733 the two rivers together with ice blocks, stones and gravel covered all the farmland. The farmers were forced to beg food in order to survive and they were therefore totally unable to pay their taxes. The court found only miserable conditions; starving people and fields covered by ice blocks, boulders, stones, and gravel. The court therefore decided (later confirmed by the authorities) that the farmers should not pay taxes for the years 1734-35. When the court visited the farm in 1743, it stated that the glacier tongue was only 60 metres from the place where the houses were located before 1728.

On 12th of December 1743, an avalanche from the glacier hit the farm houses rebuilt in 1728. All the houses, people, and domestic animals were swept away. Only a servant who was a soldier (his name was Anders Pederssen Mochleøen), a 12-year old boy, and two cows survived. The following year (1744) there was an administration of the estate that had belonged to Gullak Tungøen (the farmer) and his wife, and only the two cows, a pillow, two waistcoats, and an old sack were found. After this tragedy the farm was never rebuilt and it was deleted from the land register. About 80 years earlier Tungøyane had been one of the wealthiest farms in Oldedalen, feeding 3 horses, 40 cows and sheep.

From the historical documents, it is easy to reconstruct the natural processes that led to the catastrophe of Tungøyane. Before 1650 they "saw the glacier as a white cow on the skyline", meaning that there was only glacier ice on the Jostedalsbreen plateau above Brenndalen at that time. In the 1680s and -90s, the regenerated glacier started to damage the pastures in Brenndalen and caused floods over the farmland in Oldedalen. Around 1700 the glacier front
reached the valley mouth above the houses. This means that the glacier advanced 4.5 km in only 50 years (90 metres per year on average!). Between 1700 and 1728, the glacier flowed through the canyon behind the houses, thus threatening the houses, which in 1728 were moved to a higher terrace nearby where the farmers felt safe. On 12th December 1743, an avalanche (ice blocks, water, sand and gravel) from the glacier front resting on the rock bar above the houses, led to the final destruction of the farm.

References


Briksdalsbreen, western Norway: climatic effects on the terminal response of a temperate glacier between 1901 and 1995

Atle Nesje, Department of Geography, University of Bergen

Briksdalsbreen (11.94 km²) is a steep outlet glacier from Jostedalsbreen (Fig. 1), the largest icecap on mainland Europe. The glacier ranges in altitude from 1910 to 350 m (1560 m) over a distance of 6 km (Østrem et al., 1988). Marginal moraines in the mouth of Briksdalen (Fig. 2) are morphostratigraphically correlated to the Erdalen event (Nesje, 1992), a climatic deterioration which occurred ca. 10,000 calendar years BP. The "Little Ice Age" maximum position is marked by marginal moraines at Kleivane (Fig. 2). According to lichenometric measurements, Briksdalsbreen attained its maximum "Little Ice Age" position around A.D. 1760-65 (Pedersen, 1976). In A.D. 1870 the glacier front had retreated to the end of the cart road (Figs. 2 and 3). During the 1930s and 1940s the glacier front retreated significantly, reaching a maximum annual retreat in 1948 with 84 m (see Fig. 5a). The distal part of the proglacial lake Briksdalsbrevatnet (maximum water depth of 20 m; MacManus and Duck, 1988) was deglaciated in the early 1940s, while the minimum glacier extent was reached in the early 1950s (Figs. 4 and 5b). Between 1952 and 1973 the glacier front was more-or-less in the same position, however, between 1974 and 1980 a slight advance occurred. In 1988 a significant glacier advance started, which culminated in 1994 with 80 m, the largest annual advance recorded this century. At present, the glacier front is pushing up glacio-lacustrine sediments. In the autumn 1995 Olav Kvame, a farmer from Oldedalen, found a tree log protruding from the terminal moraine which is under formation in front of Briksdalsbreen. The tree log, which contains 62 annual rings, was identified as a willow (Salix sp.). The log was dated at the Trondheim Dating Laboratory to be about 8400 calendar years old. The Salix log is from the first part of the time interval when Jostedalsbreen is interpreted to have been melted away (see part on the Holocene glacier fluctuations in the Jostedalsbreen region).
Figure 1. The Briksdal - Briksdalsbreen area. Contour interval 50 m Scale, 1 : 50 000
Fig. 2. The Briksdalen valley and the glacier foreland of Briksdalsbreen. Marginal moraines and glacier-front positions at different times are indicated. Contour interval 25 m. From Pedersen (1976) and Nesje (unpublished).
Fig 3  Briksdalsbreen in 1870. Photograph: K. Knutsen.

Fig. 4  Briksdalsbreen in 1952. Photograph: Unknown.
Changes in summer temperature and winter precipitation 1901-95

As an indicator of summer (1 May-30 September) temperature in western Norway, records from the meteorological station in Bergen were used. To evaluate the impact of temperature on the frontal fluctuations of Briksdalsbreen, Nesje (1989) used records from the meteorological station Oppstryn at the NW side of Jostedalsbreen. The station was, however, closed in 1991. The Bergen record (Fig. 5c) shows that mean summer temperatures between 1901 and 1930 were below the 1961-90 normal (12.5 °C). The summer temperature in 1923 (10.3 °C) was the coldest measured this century. In the 1930s and 1940s, in contrast, summer temperatures were the warmest recorded this century. The warmest summer temperature was measured in 1947 (14.7 °C). From the early 1950s up to the present summer temperatures have fluctuated about the 1961-90 mean.

The winter (1 October-30 April) precipitation record from the meteorological station at Briksdal (Fig. 5d) in the vicinity of Briksdalsbreen shows that the mean winter precipitation from the turn of the last century until the late 1920s was higher than the 1961-90 normal (887 mm). The largest annual winter precipitation during this century was measured in this period (1836 mm in 1905). Between 1930 and 1960, mean winter precipitation fluctuated around the 1961-90 normal. However, the early 1960s were characterised by less winter precipitation than the 1961-90 normal. From the late 1970s the amount of winter precipitation has shown an increasing trend. The winter precipitation between 1988/89 and 1994/95 was high every year, highest in 1989/90 with 200% of the 1961-90 normal. The winter precipitation in the 1995/96 accumulation season has been significantly lower than the normal.

Frontal fluctuations and climate

Between 1901 and 1931, the front of Briksdalsbreen was in a more-or-less stable position (Fig. 5b). Warm summers in the 1930s and 40s (Fig. 5c) caused negative net balance (Fig. 6a) and a significant retreat of the glacier, reaching a maximum rate of annual retreat of 84 m in 1948 (Fig. 5a). Cooler summers in the late 1940s led to positive net balance and a stabilisation of the
Fig. 5 a) Annual glacier-front variations of Briksdalsbreen between 1901 and 1995. Data: Glaciology Section, Norwegian Water Resources and Energy Administration (NVE).
c) Mean annual summer (1 May-30 September) temperature in Bergen 1901-1995. The smoothed curve is the 5-yr running mean. Data: The Norwegian Meteorological Institute.
d) Annual winter (1 October-30 April) precipitation at Briksdal 1901-1995. The smoothed curve is the 5-yr running mean. Data: The Norwegian Meteorological Institute.
Fig. 6 a) Annual changes in mass balance on Briksdalsbreen calculated indirectly from meltwater discharge data (Pedersen, 1976). The data have been updated to 1986.

b) Annual net mass balance measurements from Nigardsbreen (data: NVE).
glacier front from 1952. Between 1988 and 1995 the terminus of Briksdalsbreen advanced 296 m, of which 220 m occurred after 1992. The lag time of frontal response of Briksdalsbreen to a change in annual net balance has been calculated to be 3-4 years, indicating that glacier termini can react rapidly to short-term temperature and precipitation changes (Nesje et al., 1995). The positive glacier net mass balance in western Norway after 1988, caused essentially by high winter precipitation, has resulted in the largest annual glacier advances measured this century: thus, Briksdalsbreen advanced 220 m between 1993 and 1995. Present knowledge about the glacier history in western Norway since the termination of the last glaciation about 9000 BP (e.g., Nesje, 1992) suggests that annual glacier expansions of such magnitude only occurred during the "Little Ice Age" (ca. A.D. 1650-1920). Historical data (e.g., Grove, 1988) show that Nigardsbreen advanced 2.8 km between 1710 and 1735, giving a mean rate of advance of 112 m yr\(^{-1}\). From the western side of Jostedalsbreen, the adjacent Brenndalsbreen advanced 4.5 km between 1650 and 1700 (average advance rate of 90 m yr\(^{-1}\)) (Nesje, 1994).

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The rockfalls in Ramnefjellet, Loen

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This review of the Loen rockfalls is mainly based on Nesdal (1983) and Grimstad and Nesdal (1991). From Ramnefjellet in Loen (Fig. 1), seven major rockfalls were released during the period 1905-1950 (Table 1), two of which (15 January 1905 and 13 September 1936) generating huge waves in the lake Lovatnet, killing 134 persons altogether. In total about 3 million m$^3$ rock masses have fallen down from Ramnefjellet since 1905.

Table 1

Observed rockfalls in Ramnefjellet, Loen.

<table>
<thead>
<tr>
<th>Date</th>
<th>Volume of rock (m$^3$)</th>
<th>Volume of till/scree (m$^3$)</th>
<th>Max. height of outfall (m)</th>
<th>Max. wave height (m)</th>
<th>No. of people killed</th>
</tr>
</thead>
<tbody>
<tr>
<td>15 January 1905</td>
<td>50,000</td>
<td>300,000</td>
<td>500</td>
<td>40,5</td>
<td>61</td>
</tr>
<tr>
<td>20 September 1905</td>
<td>ca. 15,000</td>
<td>ca. 50,000</td>
<td>400</td>
<td>&gt;15</td>
<td>0</td>
</tr>
<tr>
<td>13 September 1936</td>
<td>1 million</td>
<td></td>
<td>800</td>
<td>74</td>
<td>73</td>
</tr>
<tr>
<td>21 September 1936</td>
<td>ca. 100,000</td>
<td></td>
<td>800</td>
<td>ca. 40</td>
<td>0</td>
</tr>
<tr>
<td>6 October 1936</td>
<td></td>
<td></td>
<td></td>
<td>?</td>
<td>0</td>
</tr>
<tr>
<td>11 November 1936</td>
<td>&gt;1 million</td>
<td></td>
<td>800</td>
<td>&gt;74</td>
<td>0</td>
</tr>
<tr>
<td>22 June 1950</td>
<td>ca. 1 million</td>
<td></td>
<td>800</td>
<td>ca. 15</td>
<td>0</td>
</tr>
</tbody>
</table>

Ramnefjellet is situated on the southwestern side of Lovatnet, which is approximately 12 km long and up to 1.5 km in width (Fig. 1). In front of Ramnefjellet the lake is only 400 m wide because of the headland Nesodden on the other side. The lake level in Lovatnet is 52 m a.s.l. and the maximum water depth is 138 m (NVE, 1984). The scarp after the latest rockfall is sloping approximately 80° between 350 and 800 m a.s.l. The slope is partly stepped and the area under the rockfall scarp is sloping in average 40° down to the lake (Fig. 2). The water depth in the lake in front of Ramnefjellet has decreased after each rockfall. Therefore, the lake is at present too shallow to generate damaging waves in connection with possible future rockfalls on the same place as the previous ones.
Fig. 1. Map showing the Lovatnet area and the location of Ramnefjellet. Water depths in Lovatnet after NVE (1984). The maximum wave heights (in metres) caused by the rockfall in 1936 are indicated (adapted from Bjerrum and Jørstad, 1968).

Adapted from Reusch (1907).

The present situation.

Fig. 2. Cross-sectional sketches from Ramnefjellet, illustrating the main joint pattern (adapted from Grimstad and Nesdal, 1991).
Bedrock geology and structures

The bedrock consists of medium to fine grained granitic gneisses with veins and layers of schistose amphibolite, rich in mica and occasionally clorite and pyrite. Large amounts of water are reported to have flushed out from the boundary between the gneiss and the amphibolite in some parts of the rockfall scar. In the northern part of the scar, the foliation is dipping 40-50° in an easterly direction, while its dip is turned about 40° in the southern part. The remaining rock masses in the southern part of the rockfall area are partly resting against more stable bedrock.

Bugge (1936) described a mineral alteration in some of the schistose amphibolite layers, and supposed that the first rockfall was triggered by a collapse in an altered layer. In addition to the foliation, there are north-south striking joints dipping steeply (ca. 80°) toward the west and toward the lake (Fig. 2). This set of joints is regional and can be followed southward several kilometres up the Nesdalen valley. The spacing in these joints is probably larger than that of the foliation joints. Some of these sub-vertical joints have been reported to be open (Jørstad, 1954). Some vertical joints are also reported, two of which can be seen today. Descriptions given after the 1936 rockfalls indicate that water pressure may have been the triggering mechanism. Eyewitnesses have told about sudden outbursts of water under high pressure, like water jets, coming from certain places just above the water-bearing joint filled by alteration materials in the upper limit of the scar from 1905 (Bugge, 1936). Those water outbursts came after gushes of small black-coloured stones and gravel from the same joints. The water jet was pulsating, each pulse lasting 1-2 minutes with 6-7 minutes intervals. This high water pressure may have been connected with the ground water reservoir acting in communicating joints in the overlying mountain massif behind the rockfall scar. The partly open north-south striking sub-vertical joints are crossed by one major and some minor streams.
The 1905 rockfalls

The first disastrous slide occurred at midnight between 15 and 16 January 1905. A rockfall of about 50,000 m$^3$ fell down from about 500 altitude above the lake and released approximately 300,000 m$^3$ of till and scree deposits at the footslope. The released mass of rock was estimated to have been 10 m thick, 100 m high and 50 m wide. The rockfall left a portal-shaped, overhanging scar about 500 m a.s.l. (left side in Fig. 2). The slide generated waves with maximum height of 40.5 m above the lake level at Nesodden, opposite to Ramnefjellet. At Nesodden a steamboat was lifted vertically 17 m above the lake level and moved a horizontal distance of 250 m away from the lake shore.

At Bødal, the waves reached a maximum height of 14.5 m and killed 27 persons and completely destroyed all the farmhouses in the range of the waves. At Nesdal, the waves reached a maximum height of 15.5 m and killed all the 34 inhabitants (five farms). At the lake outlet some 8.5 km to the northwest, the waves reached a maximum height of 5.8 m and destroyed the bridge across the outlet river. All except one of the 80 boats on the lake were destroyed. In 1905 there was no road along the lake. The only access to the disaster areas in Bødal and Nesdal was a narrow, winding path, which was damaged by an avalanche on the stormy day after the disaster.

On the 20th of September 1905, about 15,000 m$^3$ of rock fell down from a maximum outfall height of 400 m and, together with about 50,000 m$^3$ of till and scree material, the masses generated waves with a maximum height of >15 m. Luckily nobody was killed.

The 1936 rockfalls

The first of the 1936 rockfalls, containing about 1 million m$^3$, was released at 5 o’clock in the morning of 13 September. The maximum wave height at Nesodden was 74.2 m (Fig. 1), 33.7 m higher than in the disastrous 1905 rockfall. The remnants of the steamboat, left 17 m above the lake in 1905, was lifted to an elevation of 50 m, some 350 m from the lake shoreline. In
Bødal, all the nine farms rebuilt on a higher level after the 1905 rockfall, were completely destroyed, and 43 persons in Bødal were killed by the waves, which reached a maximum height of 44.9 m (Fig. 1). At Nesdal 3 out of 7 farms were destroyed, and 26 persons were killed by the waves (maximum height 23.2 m). Two other farms, at Hogrenning and Vassenden, were destroyed, and two persons were killed at each of the two farms. At Vassenden, 8.5 km from Ramnesjellet, the waves reached a maximum height of 12.6 m. Originally, Nesodden was covered by till and the foothill of Ramnesjellet by till and a top layer of rock debris (talus). The waves generated by the rockfalls, however, swept most of the fertile soil away, leaving mostly bare bedrock and stony ground.

Later the same year; 21 September, 6 October, and 11 November, rockfalls occurred in Ramnesjellet, of which the latest was the greatest, with more than 1 million m$^3$. This rockfall, which came from about 800 m, generated waves with a maximum height of >74 m, but nobody was killed. The Nesdal farm was not rebuilt after the 1936 rockfalls.

The 1950 rockfall

On 22 June 1950, 1 million m$^3$ of rock fell down from about 800 m, but because the part of the lake under Ramnesjellet was mostly filled up with material from the previous rockfalls, the generated waves reached a maximum height of only about 15 m. Nobody was killed.
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**Fjærland — A Fjordside Settlement**

Ole Martin Korsen, Norwegian Glacier Museum.

**Introduction**

"Fjærland" is the district surrounding Fjærlandsfjord, a branch of Sognefjord. Fjærland has 300 inhabitants, and is part of Balestrand municipality. The area has been settled since the Viking Age. The size of the population has varied over the years. At the turn of the century there was a large emigration to America.

**Climate — Agricultural conditions.**

The climate of the Fjærland area is oceanic but nevertheless severe. Mean annual precipitation is about 1800 mm, of which most fall in autumn and winter. Winters are relatively mild, however, a one meter thick snow cover is not unusual. Mean summer temperature in Fjærland is in the range of 10-15 °C.

Fjærland is an agricultural area. Conditions are good for efficient farming, with level, easily-worked fields. The soil and growing conditions are good, especially for cattle feed production. Milk and meat production are the main activities, but the smaller farms along the fjord are mostly concerned with sheep and berries. The farms cover altogether about 1000 acres, or 4 million square metres, of cultivated land. The farms average 30 acres, or 120 000 square metres, which is fairly large by West Norway's standard.

**Tourism.**

Since the end of the 19th century, tourism has been an important livelihood. Foreigners came here to see the fjord, the mountains and glaciers, and as a result, Hotel Mundal was built in 1891. Today Fjærland Fjordstue Hotell, a campsite and cabins are also available for rent. During the 1920's and 1930's there was an increase in the number of cruise ships arriving in Fjærland. Horse and wagons would transport these tourists up to the glaciers.

**Communication.**

Until recently Fjærland could only be reached by travelling by boat up Fjærlandsfjorden. In 1986 the Fjærland tunnel to Jølster was opened by former U.S. Vice-President Walter F. Mondale, (whose family and name originated in Mundal in Fjærland). In 1994 the road continued to Sogndal, making Fjærland a central point in the county of Sogn og Fjordane.
Fig. 1 An old brochure, and the price list, from Hotel Mundal from the beginning of the century.
Fig. 2  Pictures taken before the turn of the century showing Bøyumsbreen (top) (Photo K. Knudsen) and people from Fjærland in national costume visiting the front of Bøyumsbreen glacier (bottom). (Unknown photographer)
Trade and Commerce.

"Brævasshytta" is a restaurant and souvenir shop situated by the Bøya glacier. Close to the Glacier Museum one can find a gasoline station and a campsite. Close to Mundal, the centre in Fjærland, there is a workshop producing the Fjærland shoe. In Mundal there is a bank, a post office, a telephone-box, and two stores in addition to the souvenir shop, “Vevstolen”.

Nature — Landscape.

The landscape in Fjærland has been shaped by glaciers through successive ice ages during the LAST 2.5 to 3 million years (Quaternary). Towering mountains and U-shaped valleys surround large delta areas which results from the accumulation of sediment supplied by the glacier rivers. Several thousand years old deltas formed at a time of higher sea level can now be seen as terraces in the valleys. The highest terraces in Fjærland at 110 m elevation can be seen in Supphelledalen and at the head of Tverrdalen in Bøyadalen. (See also description to Quaternary map of Fjærland.)

Jostedalsbreen National Park. Parts of Fjærland lie within Jostedalsbreen National Park. The National Park covers 1230 sq. km (475 sq. miles) including Jostedalsbreen, which is the largest glacier on the European Continent with an area 487 square km. Jostedalsbreen National Park is characterized by great variation within short distances, from fjords and lowland to mountains and glaciers. The cultural landscape in the valleys below the glacier tells about settlements since long time ago. Jostedalsbreen has been in use as a transport route for several hundred years. One of the most popular routes at the southern part of Jostedalsbreen is between Lunde and Fjærland, crossing Marabreen.

The Bøyasøyri estuary. The Bøyasøyri estuary at the head of the fjord is a protected nature reserve, due to its part in bird migration during the spring and autumn. 90 species have been observed and approximately 50 species nest in the area. The larger diving ducks, like eider and common scoter, are seen in deep water. Further in, there are golden eye and the fish eating red-breasted merganser. Grassducks, such as the mallard, and longlegged waders, like the redshank, are found in the shallow water of the fjord. The common sand piper and the oystercatcher can be watched in the tidal flat. The seagulls prefer seeking food in the agricultural areas, but make use of the outer banks of gravel to rest. The grass fields are important nesting areas for waders and gulls, and lapwing, curlew and different sparrows find their food here.

Fauna. Fjærland's rich vegetation provides for an abundant wildlife. Deer thrive on the wooded hillsides, and in the autumn and spring they graze in the meadows down by the farms. The most
commonly seen animals in addition to deer are hare, lemming, weasel, ptarmigan, heron and Golden eagle

**Flora.** Coastal and inland climates meet in Fjæerland. The coastal climate has milder winters, cooler summers and higher precipitation, whilst the inland climate has colder winters, warmer summers and lower annual precipitation. The foxglove is a coastal plant, and the monkshood is an inland plant, yet both grow in Fjæerland.

Among the plants to be seen in Fjæerland are: Blue Sow-thistle, Mountain Queen, Common Butterwort, Parsley Fern, Purple Saxifrage, Glacier Buttercup, Bluebell, Alpine Hawkweed and Wood Cranesbill.

**Fjæerland - Quaternary Geology**


Quaternary geology is the study of the youngest geological period - the Quaternary. The superficial deposits covering the bedrock in Norway is mainly from this period. The last 2-3 million years of the earth's history is called the Quaternary period. This period is characterized by great climatic changes ranging from ice ages (glacials) to much warmer periods between the ice ages (interglacials). During the ice ages, the landmasses were more or less covered by glaciers (inland ice). These eroded bedrock and carried great amounts of debris. A lot of the material was transported out to sea and deposited there, at the sea bottom.

In Europe there has been at least four ice ages. In Norway, up until now, traces of only two ice ages and one interglacial have been found (Mangerud et al 1979).

The last ice age, (Weichsel), started about 100.000 years ago. Climatic conditions changed a great deal during this ice age, which led to great variations in thickness and extension of the ice sheet. During several periods the inland ice sheet was probably partly, if not fully, gone (interstadials). The largest extent of the ice during the last ice age was found 17-21.000 years ago. Scandinavia was covered by an ice sheet which was up to 3000 m thick over the Gulf of Bothnia.
During deglaciation, the ice retreated from the coastal areas, thus the coastal areas became the first bare areas. At the same time, the ice sheet became thinner and split into valley glaciers and fjord glaciers. These quickly melted back, because of a mild climate and calving into the fjords. Shorter, colder climatic periods stopped ice melting, or led to small scale glacier growth. In this way the glacier transported material could be deposited in front of the ice edge as marginal moraines or meltwater deposits.

The most distinct marginal moraine was deposited during Younger Dryas 10-11.000 years ago. In Norway this can be traced, more or less continuously, from the Swedish border in Østfold in the south eastern part of the country (Raet), along the coast up to the Russian border in east Finnmark in northern Norway. Also, at a later stage, younger, distinct marginal moraines were deposited 9-10.000 years ago in Preboreal times. The final deglaciation of the inner, central parts of the land mass was rapid. It is assumed that the whole ice sheet had melted about 8500 years ago. Then the "warm period" arrived, with a milder climate than that of today, and all the glaciers melted away. Today's glaciers are about 2500 years old.

The weight of the ice masses pressed the crust down. As the ice melted, the crust again rose; more inland than at the coast. Because the crust is stiff, it has taken a long time to regain equilibrium. Even today the land masses rise 1mm per year. The result of the uplift is that areas that were ocean- and fjord bottom during and after the deglaciation, became dry land. The highest sea level after deglaciation is called the marine limit.

The Quaternary Map
A quaternary map of scale 1 : 50 000 is an overview of the main types of superficial deposits in an area. In the process of making a map of this scale it is sometimes necessary to generalize. Important details may be exaggerated on the map. This usually concerns the size of small deposits, drainage traces and small bedrock exposures. The lines on the map can be definite, but it will also often be seen that there is a gradual boundary between different deposits. This gradual boundary is a thin line on the map. Some exposed mountain areas are mapped by the interpretation of aerial photographs. UTM-coordinates (6 digits) from the aiding map have been used to establish the whereabouts of localities.

Superficial deposits.
The different superficial deposits are grouped by how they are deposited and the conditions under which they were deposited. Therefore, the different colours of the map indicate different geological
processes having taken place. All material carried and deposited by running water will for instance have yellow or orange colours; while material carried and deposited by ice, will have green colours. The tills are subdivided; thickness is marked by different shades of colour.

**Exposed bedrock.**
Areas of exposed bedrock are illustrated in different colours when they are of sufficient size. Areas where less than half of the bedrock is covered by superficial deposits, are illustrated as bedrock. Small deposits and deposits of vague boundaries in areas dominated by other deposits or exposed bedrock. These deposits are illustrated by symbols. In areas dominated by other deposits, the symbols are used to illustrate deposits that are either too small or too thin to be illustrated by a different colour; and for deposits that are mixed with the dominating deposits.

**Grain Size.**
Grain sizes of sorted deposits, such as water- and wind transported material, are on the map based on a visual mapping in the field. A rough estimation of the material has been done, and the dominating grain size(s) at or just beneath the surface is shown on the map.

**Thickness and Stratigraphy.**
If, in an area, there are several layers of different deposits on top of each other, the top layer’s colour is shown on the map as long as the layer’s thickness exceeds 0.5 m and the area it covers is sufficient.

**Direction of ice movement.**
Glacial striation show the direction of the ice movement. The striation is formed by material underneath the glacier eroding on the rock, forming long, backshaped roche moutonnées. These roche moutonnées have a steep distal side and a less steep proximal side.

**The Geology of the Fjærland Area**
**Bedrock and landscape.**
Fjærland lies within the north western gneiss area of Precambrian rocks, i.e. augen gneiss, feldspar gneiss and migmatite. The rocks have been metamorphosed during the Caledonian orogeny.

The landscape consists of well developed and varied glacial erosional forms. The Fjærlandsfjord is the deepest, with up to 1600 m high sides and 200-400 m deep troughs. The Supphelledalen valley (830 150) has also got 1600 m high sides. The valley bottom slopes gently at the lower end. This must have
caused the fjord, just after the deglaciation, to cover an area 6-7 km further into the valley than it does today; approximately to Mikjelstølen (851 175). The Supphelledalen valley has also got well developed thresholds and troughs in the lower end. The Bøyadalen valley, with its steeper length profile, suddenly stops in a 1500 m high wall at the end of the valley, where the Bøyabreen glacier drops steeply from the Jostedalsbreen glacier.

The side valleys are mostly hanging valleys in relation to the main valleys and the fjord. The best example is perhaps the Horpedalen valley (820 120), which ends about 100 m above the eastern part of the valley floor in Fjærland (Fig. 1). The side valleys are short and steep. They end in sides that can be several hundred metres high. Higher up, 1000-1500 m a.s.l., there are several cirques which open up in all directions. In cirques facing the north, there may be glaciers above 1250 m a.s.l.

There are several smaller plateau glaciers, west and east of the Fjærlandsdalføre valley. To the west the Jostefonniglacier (725 130) and the Troget glacier (745 165) are the greatest and to the east there is the Voggebreen/Myrdalsbreen glacier (830 975), the Bjåstadbreen glacier (820 025), the Steindalsbreen glacier (870 110) and the Svardalsbreen glacier (880 160).

Ice movement.
When the area was covered by inland ice, for instance during the Weichsel maximum, the glacier streams were probably directed by the Fjærlandsfjord in a southwest direction towards the Sognefjord (Vorren 1977, Aa 1983). Later the ice movements were directed by the local topography. In the mountain areas between the Sogndalsdalen valley (the Fremstedalen and Langedalen valleys) and the Fjærlandsfjord, there are glacial striae in several directions. This shows that the area was covered by local ice caps during the last period of deglaciation. The youngest glacial striae that exist in front of the glaciers today, can be interpreted to be traces from the growing of the glaciers during "The Little Ice Age" in the years 1300 to 1900.

Deglaciation.
The Scandinavian ice sheet and the melting of this is described in the chapter of Quaternary Geology. Even though datings are missing, it is thought that most of the Sognefjord was ice free in the Allerød period 12-11.000 years ago.

In the Younger Dryas period (10-11.000 years ago) the glaciers again advanced towards the coast. The reason was a harsh change of climate with 5-6 °C lower summer temperatures on the coast.
The climate again got warmer 10,200 years ago, and the ice sheet thinned and melted quickly back, due to calving. Carbon-14 datings of shells from the mouth of the Sogndalsfjord (Aa 1983) and mud from Aurland (Bergstrøm 1975) show ages between 9700 to 9800 years, which means that there could not have been ice left in the Sognefjord at this point. After 500 years of calving, most of the glacier edge was lying at the inner end of the fjord arms, where great deltas were formed in front of the glaciers. The Fjærlandsfjord was also probably ice free 9700 years ago.

Where the Bøyadalen and the Tverrdalen valleys meet, three km from the fjord (795 150), there is a glaciofluvial delta with an ice-contact slope. This shows that the Bøyadalsbreen glacier stopped here during the period of deglaciation. The reasons may be topographic, as there is a mountain threshold here, where the valleys meet.

Moraines found in several of the side valleys are probably from Preboreal times (10,000-9000 years ago) or the first part of Boreal times (9000-8000 years ago). Below the Anestølsvatnet lake (883 045) there is an unusually great end moraine, stretching across the bottom of the valley in a several hundred metre wide belt (Fig.5). Here, several abandoned melt-water channels (from old glacial rivers) are found.

The great end moraine shows that the glacier must have come to a stop here for some length of time and it probably also advanced again. Datings are missing, but this probably happened at least 9000 years ago. Sand and silt deposits around the Anestølsvatnet lake shows that the level of the lake must have been about 18 m above the level of today, at the time of deglaciation. Up the Frudalen valley (870 035) there are several end moraines which were formed as the ice melted back in stages towards the mountain plateau between the Sogndalsdalen valley and the Fjærlandsfjord. An unusually pointed and nice end moraine in the Tverrdalen valley (885 083) dams the Nedre Tverrdalsvatnet lake. This end moraine marks a stop during deglaciation about 9000 years ago. Between Nedre (lower) and Store (great) Tverrdalsvatnet lakes (894 093) there are several distinct lateral moraines to be found. From Boreal times, continuing into the warmer periods, all the glaciers probably totally melted.

"The Little Ice Age".
During "The Little Ice Age" (year 1300-1900), the glaciers expanded until about 1750. At this point the Jostedalsbreen glacier and the smaller glaciers within the area of this map, covered the greatest area in historic times. The 1750-moraine is known by the marked difference in vegetation outside and within the moraine ridges. In the north western corner of the map, the Troget glacier (745 170) has advanced, down towards the Femtevatnet lake (735 185). The Marabreen glacier (775 206) also
expanded; in a south easterly direction towards the Trollvatnet lake. Distinct side moraines between the Sognsdalsdalen valley and the Fjærlandsfjord show that the glaciers in this area also expanded during "The Little Ice Age".

**Uplift - Marine Limit.**

At the mouth of the Horpedalen (815 122) and Tverrdalen (790 150) valleys, and 5 km into the Supphelledalen valley (848 171), the surfaces of glaciofluvial deposits probably represent marine limit. The first two places are described by Mundal (1953), who found levels at 101 and 118 m a.s.l. at the mouth of the Horpedalen valley. The uppermost surface is by Mundal interpreted to be an abrasional surface, and marine limit for this area. But, as he states, the surface is uneven and covered by erosional traces and might therefore as well be a glaciofluvial erosional level.

The highest terrace of the glaciofluvial delta at the mouth of the Tverrdalen Valley in the Bøyadalen valley, is 110 m a.s.l. Below Mikjelstølen in the Supphelledalen valley, there is a narrow, horizontal lateral terrace at the same level, where the fjord seems to have extended 5 km past the mouth of the valley. This surface in the Supphelledalen valley almost certainly represent marine limit in Fjærland and the marine limit in Fjærland is therefore interpreted to be 110 m a.s.l.

**The Bøyabreen glacier.**

In the Bøyadalen valley several moraine ridges show the history of the glacier after 1750. The oldest of these ridges, from ca.1750, is found just below Bøyastølen (797 178). It is 8-10 m high to the west of the road and 10-15 m high east of the river. Several of the younger moraine ridges have been damaged by avalanches above Bøyastølen. Where the river Jakobbakka (799 182) crosses the road, there are 6 moraine ridges, 2-3 m high. Further in, towards Brævasshytta (803 188), about 10 moraine ridges are found; the largest one by Brævasshytta is about 5 m high. The younger ridges close to the mountainsides are small, about 1 m high, and consisting mainly of boulders.

From 1750 the glacier history of the Bøyadalen valley is one of the best documented areas around the entire Jostedalsbreen glacier; through glacier measurements and photographs. The first measurements are from 1868 (S.A.Sexe) and the first photographs from 1867-1868 (de Seue). By studying photographs, Rekstad (1900) discovered that the glacier had expanded in the period 1868-72, melted back until about 1880, expanded again until 1888 and melted back until 1899. In 1899 crosses were cut into two granite blocks, one on either side of the glacier. The cross on the east side is still there today (Mjanger and Hofsøy 1989), but the cross on the west side has not been found, it might have been erased by an avalanche. Good photographs exist from the period 1872-90, taken by K. Knutsen, from
1899-1906 taken by J. Rekstad, and later taken by K. Fægri. During the period 1900-47 continuous measurements were made by K. Fægri. This shows that since the Bøyabreen glacier filled the whole lake Bøyavatnet and the great moraine ridge was formed by Brævasshytta in 1930, the glacier has continuously melted back.

In addition to historic documentation, the moraine ridges have now been dated using lichen (Mjanger and Hofsøy 1989). The results that are shown in Fig. 3 match historic data well.

**The Supphellebreen glacier.**

The Supphellebreen glacier consists of a continuous arm down to about 720 m a.s.l. The glacier tongue calves from here, and below a new glacier forms. The front of this glacier lies at 51 m a.s.l, the lowest in southern Norway today. Below the regenerated glacier, several moraine ridges show the position of the glacier front in earlier times. The oldest, the 1750-moraine, lies a couple of hundred metres above the cultivated land at Supphella (838 159). Below this, there is a sandur containing a large amount of block material. The moraine ridge within the open grazing land at Heimastølen (839 159) is 2-3 m high and has got a large content of block material. This ridge is from around 1800. The next ridge, 60 m further in, has a distinct, pointed form and is about 1.5 m high.

Because the area is easily accessible, photographic material from every decade since 1860 is found. In this way the position of the glacier tongue is known at all times in this period. The Fjellstølen hut (839 165) lies close to the 1930-moraine. There is a small sandur in front of the glacier edge.

**The Flatbre moraine.**

In the area between Flatbrehytta (826 128) and the south western edge of the glacier, there are several parallel side- and end moraines. The oldest, from 1750, can be followed from Flatbrehytta in a southeasterly direction. A younger moraine ridge lies a little further in from the 1750-moraine.

The by far largest moraine ridge is 30 m high and very pointed (Fig. 6). This is called the 1900-moraine, but also today a large part of the west side and the southwest front of the glacier lie next to the same moraine ridge.

While the southeastern part of the glacier front calves into the Supphelledalen valley, the southwestern part has been stationary through several decades. This probably also explains the size of the moraine ridge. The moraine ridge dams a lake which is in contact with the western edge of the glacier. During
heavy rainfall and melting on the glacier the water rises and runs as a large river through a 8 m deep channel in the moraine. The river runs down through the Supphelledalen valley and it may threaten the land of the Øygarden and Supphellen farms (A. Øygard pers. stat.).

*Superficial deposits.*

**Moraine.** Continuous covers of till are usually found in the side valleys. When the glaciers were thick enough they were able to move across or partly across side valleys - these side valleys would then act as “moraine traps”. The Sogndalsdalen valley and its side valleys are good examples of this. The till often contains a lot of boulder size material at the surface. A 4 m high moraine section by Vatnasete (891 017) shows a sandy till, rich in boulders.

One of the largest moraine ridges is the 15-20 m high mid moraine (878 037) where the Frudalen and Sogndalsdalen valleys meet. This can be traced several hundred metres up towards Oksli. The hill that dams the Anestølsvatnet lake (882 045) is an unusually large end moraine which stretches across the valley bottom in a several hundred metres wide belt. Seismic measurements (Fig. 5) show 70 m of superficial deposits here. This material consists of 40-50 m of water saturated moraine on the bottom, then 10-15 m of dry moraine or coarse sand and gravel with velocities ranging from 900-1200 m/s, and at the top a 4-5 m thick layer of looser packed material with velocities of 450-700 m/s. The river has eroded 20-25 m down.

*Glaciofluvial deposits.*

**Delta.** Glaciofluvial material is found as remains of deltas at the mouth of most side valleys. In the Mundalsdalen valley (783 101), remains of terraces are found up to 58 m a.s.l. At the top there is only a 12 m long backshaped erosional remains of layered gravel, sand and fine grained sand. Larger terraces are found 55 m a.s.l, south of the river and 50 m a.s.l north of the river. The largest glaciofluvial delta is deposited from the mouth of the Tverrdalen valley (793 150) and down the Bøyadalen valley. The south eastern part of the delta is deposited from the Bøyadalen valley. The highest lying terrace remains on the north side of the Tverrdalselva river, falls towards the north east and has an outer edge at 80 m a.s.l (Mundal 1953). The material consists of rounded boulders, stones and gravel, and some silt in the southernmost terraces at 24 m elevation. In the Horpedalen valley (813 122) there is a delta of steep inclined layers, of coarse gravel and sand, which covers the area from approximately 100 m a.s.l., down to the young river plain of the Bøyaelvi river.

**Glaciofluvial plains.** The largest glaciofluvial plain or sandur is found below the 1750-moraine in the Supphelledalen valley. It is approximately 300 m wide and can be traced 1 km down the valley, where
it merges with the large fan (830 154) from Krokgilja. The plains between the moraines in the
Bøyadalen and Supphelledalen valleys consist of glaciofluvial material that may range in grain size
from stones to sand.

River deposits

The Sogndalsdalen, Supphelledalen and Mundalsdalen valleys are filled with river deposits. The grain
size ranges from gravel to cobbles and boulders. The great delta reaching from the end of the
Fjærlandsfjord to the Bøyadalen and Mundalsdalen valleys, could partly be classified as a glaciofluvial
delta. Especially the Supphelledalselvi river carries glacier mud. The Bøyaelvi river carries less glacier
mud because it is partly deposited in the Breavatnet lake. But the distance from the glaciers to the fjord
is of such a distance that the classification fluvial delta has been chosen. Drilling on the Glacier
Museum grounds (807 122), show grain sizes ranging from fine sand to sand and gravel. Some thin
layers of silt have also been found. It has been told that 120 years ago, small boats were able to harbour
300 m inside of today's deltafront. This means that the delta plain expands by 2.5 m every year. Today
the embankment far into the fjord has totally changed the natural pattern of delta growth.

Rock fall and rapid mass-movement

Rock falls are fan shaped or continuous slopes below steep mountainsides. The material is sorted; the
fine material being at the top towards the apex. Rapid mass-movement deposits are the most common
deposits on the map. Snow avalanches are the most common within this group. The material is not
sorted and the cobbles and boulders have sharp edges. Steep and high mountainsides ending abruptly in
a flat valley bottom are common. The east side of the Supphelledalen valley is a good example. The
Dauremåla avalanche across the valley from the Supphella farms, has a total height of 1500 m. The
very steep avalanche path reaches up to approximately 900 m a.s.l. Above this, there is a 3-400 m high
bowl shaped accumulation area. The avalanche wind can make damage 500 m up the opposite
mountainside. In the summertime there are often rock falls and small stone falls in this area.

Rock avalanches by the Fjærlandsfjord. At the southernmost tip of the Bøyaøyri plain, by the eastern
mountainside (806 115), there are some distinct, large ridges and mounds on the elsewhere very flat
delta plain. Mundal (1953) discussed several alternative explanations as to how the ridges and mounds
had been formed. Considering the outer form and the material's composition, this can be interpreted as
being the remains of an unusually large avalanche originating in the eastern mountainside. The
avalanche is 600 m wide and can be traced from the eastern mountainside to the Bøyaelva river 600 m
further west. The outer boundaries have the shapes of horse shoes with large ridges to the north and to
the south. These fall from the mountainside towards the west, where they disappear under the surface.
The northernmost ridge is 44 m high, and has a 100 m flat area at the top. Between the ridges there are three marked cones, the largest is 30 m high. Closer to the river, the area between the avalanche ridges becomes a slightly sloping river plain consisting of fine sand and silt down to 12 m, discovered by samples from drilling (Statens Vegvesen 1990). The surface is characterized by large boulders, with diameters exceeding 6 m. In the road cutting ordinary avalanche material can be seen; i.e. unsorted masses of material with sharp edged boulders and cobbles.

From the morphology it is evident that the material was deposited in the latter part of deglaciation in Fjærland, approximately 9700 years ago, while ice remains till covered parts of the valley bottom. The steep, north eastern side of the avalanche resembles an ice-contact side. The interpretation of the three cones in the middle of the tongue of the avalanche is uncertain. They may have been supported by ice and developed their cone form as the ice melted away. The flattening of the top of the ridges is a result of the sea washing over them at levels of 30 and 40 m a.s.l, during the uplift. We know that there was large activity when it comes to avalanches in the time just after deglaciation. The mountainsides were unstable at this point because of pressure release.

*Quaternary protect-worthy deposits and areas.*

Quaternary protect-worthy deposits and areas within this map are described in "Kvartærgeologisk verneverdige førekomster og område i Sogn og Fjordane" (Aa, R. A. and Sønstegaard, E. 1994). The Flatbre and Tverrdalen moraines are now within the boundaries of the Jostedalsbreen National Park, whereas the marginal moraines in the Bøyadalen and the Supphelledalen valleys are not. These are for this reason in the need of their own management plan. The avalanche on the Bøyaøyri plain lies within the area of the new road from Fjærland to Sogndal. A plan of adjusting landscape has been made. The same applies to the area around the Frudalen valley.

Captions to figures in the Fjærland Geological Map

*Fig. 1.* The Horpedalen valley; a hanging side valley in Fjærland. Photo: A. R. Aa 1988.

*Fig. 2.* Glacial troughs and sediments of the Fjærlandsfjord. Modified after Vangsnes 1981.

*Fig. 3.* The age of the moraine ridges in the Bøyadalen valley (800 180). After Mjanger and Hofsøy 1989.

*Fig. 4.* Profiles of the lower Supphellebreen glacier in the period 1750 - 1965. After Orheim 1970.

*Fig. 5.* Seismic refraction profiles of the deposits below the Anestølsvatnet lake (886 034). After Kjølseth 1968.

Legend:  
Surface  
Layer boundary  
Assumed bedrock boundary  
Seismic velocity
Fig. 6. "The 1900-moraine" between the Flatbrehytta tourist hut (827 178) and the south western edge of the Supphellebreen glacier is approximately 30 m high, and one of the largest moraine ridges in historic times in Norway. Photo: A. R. Aa.

Glaciology of Fjærland

Fjærland is surrounded by glaciers with a great variation in size; from small cirque glaciers to Jostedalsbreen, which is the largest glacier on the continent of Europe. The ice field stretches about 80 km from northeast to southwest, and cover an area of 487 km². Drillings and radioecho soundings have shown that Jostedalsbreen is up to 600 m thick. Highest elevation is at 1957 m (Høgste Breakulen), and lowest elevation is the foot of Supphellebreen at 60 m. More than 30 named outlet glaciers reach down into the low-lying valleys. Two of these outlets, Supphellebreen and Bøyabreen, come down in Fjærland.
Supphellebreen

Supphellebreen is a 0.1 km² regenerated glacier extending from 320 to 60 m a.s.l., making this the lowest lying glacier in Southern Norway. The glacier is nourished by ice breaking off from the 50 m high front of Upper Supphellebreen (called "Flatbreen" locally). Flatbreen covers 11.8 km² and extends from 1740 to 720 m elevation.

Investigations of Supphellebreen and Flatbreen made by Orheim (1970) through the years from 1963 to 1967, showed that about 2 million tons of ice break off from Flatbreen every year. This was equal to adding a 15 m thick ice layer to the lower glacier, or expressed in other terms: if the ice was in drink-size ice cubes and placed end to end, they would stretch 50 times round the earth! The flow rate at the front of Flatbreen is about 2 m/day, and the response time of changes in throughflow at the snout of Flatbreen to changes in its mass balance is 2 to 3 years (Orheim, 1970).
Supphellebreen reached its maximum post-glacial extent around 1750 A.D., about 800 m further downvalley than today. Supphellebreen was even at that time a regenerated glacier (Orheim, 1970). Positions of Supphellebreen were measured by Rekstad from 1901 to 1958, and from 1964 to the present. Since 1992 the position has been measured by the Glacier Museum. The measurements show an advance of 40-50 m a year except for last year; then there were no advance at all.

**Flatbreen**

The following is taken from a chapter about Flatbreen written by Orheim (in: Aarseth, I. et al. 1988):

"The mass balance studies from 1963 to 1967. Orheim (1970, Fig. 17) gives the mass balance of the glacier in an equilibrium year.

The mass balance studies from 1963-67 were used to establish a skeleton measuring programme, selecting those stakes which were relatively easy to maintain and which correlated well with the mean annual balance. Orheim (1970, fig. 18) gave the relationship between selected stakes around 1260 m elevation and the mean balance. This elevation was close to the multi-annual mean equilibrium line, and the stakes represent a large proportion of the glacier area. One-three such stakes have been measured from 1968 to the present, and from this the annual net balance has been calculated (Fig. 4.). Data are tentative for five years, and generally this measuring programme is of less precision than the 1963-67 studies. However, the annual variations show high correlation with other nearby glaciers, such as Nigardsbreen and Hardangerjøkulen. Possibly the relationship given by Orheim (1970) is in error by 0.1-0.2 m, as the cumulative balance curve indicates a net gain of 9.20 m water equivalent, or +0.37 m w.e./year. This seems high in view of the fact the front of Supphellebreen has only advanced a few m between 1968 and 1988."

**Bøyabreen**

Bøyabreen is 5.7 km long and has an area of 13.9 km². The highest mountains on either side of the glacier are at 1500 m elevation. The skyline of the glacier is at 1000 m elevation, and here the glacier is probably ~100 m thick. The glacier stretches down to 490 m elevation, where the ice break off. Below the snout a regenerated glacier is located between 300 m and 150 m elevation.
Due to the glacier advance the last years, the regenerated glacier is now for the first time since the end of the 1940's connected to the main glacier outlet on the right hand side. To document the glacier advance, the Norwegian Glacier Museum is monitoring Bøyabreen with a time-lapse video camera, mounted inside Brævasshytta.

In the Bøyadalen valley several moraine ridges show the history of the glacier after 1750. The moraine ridges have been dated using lichen (see description to Quaternary map).

One should not walk closer than 1 km distance to the glacier because of rock- and ice avalanche danger.

*Fig. 5  Bøyabreen at the turn of the century*
Fjærlandsfjord

Fjærlandsfjord is a northgoing branch of Sognefjord, which is the largest fjord system in Norway, penetrating about 200 km inland from the coast. Sognefjord is formed by a combination of exploitation of rock structure and subaerial and subglacial processes. The fjord system follows zones of rock-structural weakness (Nesje & Whillans 1994). Nesje & Sulebak (1994) calculated the average rate of glacial erosion in the Sognefjord drainage basin to about 0.4 mm yr⁻¹, assuming glacial erosion in a period of 1 million years during the past 2.57 million years.

The shape of Fjærlandsfjord alternates between deep troughs and shallower thresholds. The depth
Fig. 7  Longitudinal profile of Sognefjord. Summit levels and hanging tributary fjord bottoms are shown. (From Nesje & Whillans 1994.)

Fig. 8  Contour map and longitudinal profile of the sediment distribution of Inner Fjaerlandsfjord. (From Aarseth et al 1988 - modified from Vangsnes 1981.)
varies from about 110 m near Mundal at the inner part of the fjord, to about 300 m at the outer part near Hella. Fjærlandsfjord, as well as nearly all the tributary fjords “hang” above the main fjord (Holtedahl, 1967). Sognefjord is about 1000 m deep outside Hella (fig. 7).

On the sea floor of the fjord there are layers with sediments on top of the bedrock. This is material deposited by glaciers and rivers, and by avalanches down the mountain sides (Aarseth et al 1989). The thickest layers are almost 200 m thick. These are located in the outer part of the fjord, and were deposited by the glacier. After the retreat of the glacier – about 9700 years ago – the sediments have mainly been rock flour from the glacier rivers. Even though the colour of fjord in the summer is grey due to glacier flour, the sedimentation is only a couple of mm per year on the sea floor. By Mundal the sedimentation is about 4 mm, and 1 mm a year 1 km outside Distad (Vangsnes 1981).

Fig. 9 Fjærland and the inner part of Fjælandsfjorden (Photo E. Distad.).
The Norwegian Glacier Museum

is a non-profit foundation established by the International Glaciological Society, Norwegian Mountain Touring Association, Norwegian Polar Research Institute, Norwegian Water Resources and Energy Administration, Sogn og Fjordane Regional College, The University of Bergen and the University of Oslo.

The aim of the Norwegian Glacier Centre is to collect, create and disseminate knowledge of snow, ice and glaciers. This includes understanding both the natural environment and the interaction between mankind and nature. The glacier centre will stimulate the curiosity of those who know little about glaciers and inform those who know much.

New exhibit 1996:

A temporary exhibit in 1996 deals with the dramatic glacier growth in western Norway we are witnessing today. There is focus on four of the same glaciers as we will visit during the conference excursions; Brigsdalsbreen, Bøyabreen, Supphellebreen and Nigardsbreen. In the models, moraine ridges indicate the limits of glacier extension in times past. A transparent part of the glacier snout shows the growth of the glacier since the time of least expansion up until 1995. An 8,400 year old piece of tree log found in the end moraine in front of Brigsdalsbreen is also exhibited.

Fig. 10 The Glacier Museum in Fjærland. (Photo Sogn Dagblad.)
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Glacial geology of the Sogndal-Jostedalsbreen area, western Norway
Asbjørn Rune Aa, Sogn og Fjordane College, Sogndal

Bedrock and topography
The bedrock of this area consists of three main units (Bryhni et. al. 1986):
1. Precambrian basement comprising band gneisses, augen gneisses and granites.
2. The cambro-silurian phyllite and mica schists with bodies of serpentine.
3. The caledonian nappe with gneisses, gabbro, amphibolite and anorthosite.

Mountains, fjords, ice eroded valleys, hanging tributary valleys, and cirques are common landforms. In Lusterfjorden, a tributary fjord to Sognefjorden, there is a 650 m deep rock trough at the interfluve of the earlier Lusterfjord glacier and the Veitastrond glacier. The mountains on both sides reach altitudes of 1100 m. «Over-deepened» glacial troughs are common features also in Sogndalsfjorden and Veitastrondsvatnet. The troughs are separated by rock bars, a total number of 11 rock bars are found from Sogndalsfjorden to Austerdalsbreen (Fig. 1). Downcutting into rock bars by running water is a striking feature of which a good example, the «Årøy canyon» can be seen between Sogndalsfjorden and Hafslovatnet, and from Hafslovatnet to Veitastrondsvatnet there is a deep canyon with several large potholes. Glacial cirques are most widespread in the area between Veitastrond and Fjærland.

The north-east and south-west running fjords and valleys coincide with the main caledonian structures, while a north south fracture system gives origin to some of the main valleys.

Ice movements
Four glaciation phases have been reconstructed on the basis of glacial striae (Aa 1982a). The oldest is characterised by south-moving ice streams from the Jostedal glacier plateau and Jotunheimen. These oldest ice streams flowed towards the SW and SSW in the southern areas towards Sognefjorden (Fig. 2). The SSW ice movements of phase 1 veered towards the SW during phase 2, caused by a more easterly position of the ice divide. It is likely that phase 2
Fig. 1  Rock bars in Austerdalen and Veitastrondsvatn (King 1959).
Fig. 2  A reconstruction of the ice movements in 4 phases in the Sogndal-Jostedalsbreen area (Aa 1982a).
corresponds to the Late Weichselian maximum. The regional ice movement towards the SE during phase 3 suggest ice flow from the Jostedal Plateau, probably of late Younger Dryas-Preboreal age. Glacial striae from this time have been found as far east as 8 km east of Sogndal. During the actual deglaciation local ice caps existed in the mountains (phase 4).

Deglaciation

During early Preboreal Chronozone the ice front in the Sognefjord area retreated rapidly from the coast to the heads of the fjords (Fig. 3). A 14C-dating obtained from marine shells at the mouth of Sogndalsfjorden (Fig. 3), 9 740+/-120 B.P. gives a minimum age for the retreat of the glacier from the Sogndalsfjord area.

At the heads of the fjords the ice front stopped temporarily due to reduced calving, building up frontal deposits. A number of these deposits can be correlated with distinctive lateral moraines, indicating glacial readvances. Nice examples are found at the mouth of Jostedalen where the Gaupne and Høgemoen deltas (Fig. 4) with corresponding lateral moraines along Jostedalen valley, mark two Preboreal halts or readvances during the general deglaciation. The Høgemoen event is at least partly due to climatic deterioration (Vorren 1973).

The glaciofluvial frontal deposits at Årøy and Solvorn (Fig. 4) was probably formed simultaneously, corresponding with lateral moraines on both sides of Hafslovatnet and Veitastrondsvatnet. These marginal deposits mark a Preboreal halt and probable readvance called the Årøy-Solvorn Event (Aa 1982a). The younger Hafslo event (Fig. 4) is documented by glaciofluvial frontal deposits in the western part of Hafslo-vatnet, and corresponding lateral moraines along Veitastrondsvatnet (Aa 1982a,b, Aa & Nesje 1986).

During deglaciation the fjord level was 130 m higher than at present in Sogndal, 119 m in Årøy at the head of Sognefjorden, and 99 m in Gaupne, indicating a delayed deglaciation in the Veitastrond and Jostedalen areas.
Younger Dryas
The Ra-end-moraines
Ice flow direction
Preboreal
Approximate position of the ice margin during the Eidfjord-Osa/Loven/Gaupne/
Åroya-Solvorn Events
Ice flow direction
Ice flow direction, tentative
Approximate position of local ice caps

FIG. 3
Reconstruction of extent and flow directions of the glaciers during the Younger Dryas and the Preboreal Chronozone in the Sognfjord area.

after Verben (1973), Bergstrøm (1975) and Aa (1992a).
Fig. 4  Glacial map of the Hafslø-Veitastrond area (Aa 1982a).
The drainage history of the Hafslo basin

During the deglaciation the river which now has its outlet at Årøya drained to Solvorn (Fig. 4) for some time. Great quantities of glaciofluvial/glaciomarine meltwater sediments were deposited in Lusterfjorden at Solvorn. At first, the meltwater drained through a channel 265 m a.s.l. in the valley side to the east of Hafslovatnet. Then a lower pass-point 200 m a.s.l. at Galden (road junction) melted free. At this time there must have been an ice-dammed lake to the NE of Hafslovatnet where glaciolacustrine sediments were deposited. Later the pass-point at Årøya 169 m a.s.l. was deglaciated and the river found its present course. This outlet, too, is a canyon containing several potholes.

The altitudinal difference in marine limits of Solvorn at Lusterfjorden and Årøya at the head of Sogndalsfjorden, 123 and 119 m a.s.l., respectively, is an indication of earlier deglaciation of Lusterfjorden than Sogndalsfjorden.

Avalanches along Veitastrondsvatnet

Not surprisingly, the high and steep fjord and valley sides are exposed to avalanches of different kinds. Avalanche activity increases towards Jostedalsbreen together with the increase in precipitation. The steep short valleys, especially on the eastern side of Veitastrondsvatnet make effective traps for wind-transported snow. The road from Hafslo to Veitastrond is less than 40 years old but has been closed for five winters by snow avalanches. The people of Veitastrond have naturally become to being isolated 3-4 months during winter time. In 1979, one of the most active «avalanche-years», a school-bus was closed in between two avalanches, and was not released until spring time. The safety is better at present, as several tunnels have been built in the last 5 years.

Meltwater deposits in the Veitastrond - Austerdalsbreen area

The 4,5 km long delta in Veitastrondsvatnet has been built up since deglaciation 9000 (?) yr B.P. Meltwater from Langedalsbreen and Austerdalsbreen are the most important sources. In this delta the coarser grains have been separated along a stable river channel, and the main part of the delta plain consists of finer sediments. There are only two deltas of this kind
Fig. 5 Austerdalsbreen moraine-ridge sequence. Moraine ridges are lettered (for the foreland as a whole), and numbered (for each side of the foreland separately) as in Table 3. Also shown are historical dates for moraines (dates in brackets indicate an element of uncertainty) and the lichen-measurement plots used for dating purpose. From Bickerton and Matthews 1993.
around Jostedalsbreen (Bogen 1981). The other one is in Leirdalen, a tributary valley to Jostedalen, and is at present «drowned» by water backe up behind the Leirdalen/Tunsbergdalen hydroelectrical dam.

The lower part of the Veitastrond delta has been flooded several times in recent years. Notice the high ground floor-walls of living houses in Veitastrond.

At the junction of Langedalen and Austerdalen the valley bottom is filled by a coarse-grained sandur, in which numerous braided stream channels, at present dry, can be seen. The sandur downstream from Tungestølen was active in the 1930's. Photos from this time show flowing rivers over the entire sandur plain.

The «Little Ice Age» moraines of Austerdalsbreen

The moraine sequence of Austerdalsbreen has been studied by a number of authors, from Rekstad (1911) to Bickerton and Matthews (1993). According to Bickerton and Matthews (op. cit.), who used the lichenometric method, at least 15 moraine-ridge complexes (A-O in Fig. 5) can be distinguished. Extensive areas of glaciofluvial outwash have eroded some terminal moraine ridges and steep valley sides have restricted the development of lateral moraines. Nevertheless, most ridges are of sufficient length to allow establishment of at least five plots for lichen measurement. Considerable avalanche, rockfall and rockslide activity has affected some areas, most notably on the western side of the foreland. Since the deposition of the youngest moraines in Fig. 5, Austerdalsbreen has retreated behind an extensive rock bar (Bickerton and Matthews op. cit.).

No moraine ridge can be classified as being of exact historical age, as no measurements of the position of the glacier snout were made between 1920 and 1932, and Rekstads mark for recording the annual position of the glacier has not been found. However, moraines N and O (median predicted dates 1926 and 1932, respectively) were deposited between 1920 and 1932, and the significant advance of 62 m measured between 1905 and 1909 almost certainly produced moraine M (median predicted date 1906). Lichenometric dating results using
Fig. 6  Map of Austerdalsbreen and its foreland (King 1959).
Rhizocarpon subgenus are summarised for each moraine ridge in Table 1. (From Bickerton and Matthews, op. cit.)

The front position of Austerdalsbeen has been stable since the 1970's, however, the tongue has grown higher at its margins during the last two years, and the front is also expected to start advancing soon.

The ice tongue of Austerdalsbreen

Austerdalsbreen (Fig. 6) is an outlet glacier from the Jostedalsbreen ice cap, fed by two 800 m long icefalls, Odinsbre and Torsbre. The westernmost icefall, Lokebre is not connected with Austerdalsbreen at present. The ice velocity is 2000 m·a⁻¹ in the upper parts of the ice-falls, slowing down to 50 and 13 m·a⁻¹ at 2 km and 20 m from the terminus, respectively. (Fig 7; Glen 1961, King & Lewis 1961).

On the flat ice tongue downglacier from the ice falls are the well known ogives, alternating

<table>
<thead>
<tr>
<th>Ridge</th>
<th>Predicted size (mm)</th>
<th>Predicted date (AD)</th>
<th>Lichen size (mm)</th>
<th>Lichen date (AD)</th>
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<td></td>
<td></td>
<td></td>
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<tr>
<td>O</td>
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</tr>
<tr>
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<td>82³</td>
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<td>116.0</td>
<td>1764</td>
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<td>1762</td>
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</table>
Fig. 7  Variation in the rate of flow of Austerdalsbreen, from the head of the glacier (ice-fall) to its snout (King and Lewis 1961).
light and dark ice-bands. Ogives are thought to be formed in the ice falls, the light, air-filled ice during winter time, and the dark ones during the summer (Fig. 8). There are a total of sixty pairs of light and dark bands from the ice falls down to the front of the glacier.

Austerdalsbreen is one of the classic field sites for modern glaciology. The ogives were examined by King and Lewis (1961), the glacier movement by Finsterwalder (1951), Glen (1956, 1958a,b, 1961), Nye (1959) and Glen and Lewis (1961). Bull and Hardy (1956) examined the thickness of the glacier tongue by gravimetric measurements. Ward (1961) cored the ice tongue. Eyles and Rogerson (1978) examined the medial moraine of the glacier tongue. Front position measurements have been made Norsk Polarinstittutt since 1905, apart from the period 1918-32.

Glen (1956) measured deformation of ice in a 30 m long tunnel (Fig. 9) which was dug horizontally into the ice at the foot of Odins-bre icefall. He showed that waves, the ogives form across the glacier in this region. The tunnel was decreasing in length due to compression
Fig. 9 Velocity vectors for the stakes on the surface of the glacier (A) and for the pegs in the tunnel compared with surface stake 4 (B) (Glen 1956).

In A the arrows join the position in space occupied on 6, 19, Aug. and 1. Sept. 1955 by ice which was at the surface of the glacier at the stakes on 1. Sept. The glacier profile drawn in the vertical section is the approximate position on 1. Sept.

B shows velocity vectors for the pegs in the tunnel and surface stake 4, relative to the tunnel entrance stake. A rotation has been subtracted from all vectors to make the relative velocity of the 125 ft. Peg horizontal. The stress system deduced in the text from these relative velocities is shown in the centre.
strain along the tunnel axis, and the tunnel was closing far more rapidly than would be predicted from the weight of overlying ice alone. Measurements of surface stakes confirmed the observations of compressive stress acting in the glacier in a longitudinal direction (Glen op. cit). The compressive stress was estimated to be about 3 bars. The fine bands observed on the walls of the tunnel were within 10 degrees of being perpendicular to the maximum compressive stress.

Glen (1958a) measured the flow velocity at two places where clear ice was moving over unfragmented rock. It was found that the ice was sliding at a velocity of the same order of magnitude as the velocity of flow near the center of the glacier in this region. The large amount of slip is taken as evidence for a high slip rate all along the ice-rock interface. A number of stakes from the Cambridge Austerdalsbre Expedition 40 years ago can still be found on the Austerdalsbreen glacier tongue.
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The 1979 flood in Jostedalen
Nils Haakensen, Norwegian Water Resources and Energy Administration.

Introduction
The Jostedalen Valley is a narrow valley with a relatively thin soil cover so that large areas of the valley consist of exposed rock. This creates the potential for a rapid runoff. Accordingly, the floods are usually large but short-lasting. However, parts of the Jostedalsbreen icecap and other minor glaciers (in total 232 km² glacier area) drain into Jostedalen (Østrem & al. 1988). These glacier areas constitute a considerable melting potential in warm periods.

On August 15th 1979 an extraordinary large flash flood ravaged Jostedalen Valley. The flood was caused by extreme precipitation combined with snow melting from glacierized areas of the Jostedalsbreen icecap and some other adjacent minor glaciers. This flood exceeded the former highest observed flood, which occurred on August 15th 1898. In the paper “Søgningen” from 1898 the following can be quoted:

The river Jostedalselven was extremely large on 13. and 14. August. Old people have not seen it that large before. It caused severe damage for people living close to the river. Fields and meadows were destroyed. Four bridges were totally destroyed. The flood culminated on the August 15th. It is the largest ever known in Jostedalen.

Exactly 81 years later, on August 15th 1979, a new flood struck Jostedalen. This flood was caused by extreme precipitation combined with snow melting from the glacierized areas. The flood considerably exceeded the 1898-flood, and the damages were enormous.

Weather Conditions
The western part of southern Norway on August 13th was dominated by a warm and humid southerly air current. The air was unstable and minor low pressure areas were formed. Air temperature was relatively high. During the following night, precipitation increased. In Jostedalen there developed a strong convergence zone which intensified the precipitation and the wind. On the August 14th temperature increased strongly, up to 17-19°C near the ground. Some föhn-effect resulted in a local maximum temperature of 23°C. Due to the convergence zone over Jostedalen wind speed increased and this combined with the high temperature, produced an accelerated snow melt from the glacierized areas (Gjessing & Wold, 1980).
Figure 1. The Jostedalen area. Contour interval 50 m

1) Nigardsvann recording water gauge. 2) Myklemyr recording water gauge. 3) Leirdøla with lake Tunsbergdalsvann hydropower reservoir.
Fig. 2  A precipitation map showing the distribution of rainfall in 24 hours (from 0700 the Aug. 14th to 0700 the next day). Most of the rain fell between 0700 and 1900 the Aug. 14th. (From Gjessing & Wold 1980)
Fig. 3  The river from Nigarsvann took a new course and broke the road. An old mill was totally smashed. The millstone is visible by the waters edge. (Photo E. Roland.)

Fig. 4  The river Sagarøyelvi transported large amounts of material, even big boulders. It was deposited in the valley near the confluence with Jostedøla where the river gradient decreased and filled up the water bed. Consequently the river broke through the river wall and carried huge amounts of stones and boulders that made great damage on houses and cultivated land. (Photo E. Roland.)
Fig. 2 shows how the precipitation was concentrated to the east side of Jostedalsbreen and the upper part of Jostedalen. Up to 80 mm precipitation was recorded in 24 hours, whereas at the inner part of Sognefjord, only 40 km away, the precipitation was less than 10 mm.

The decisive factors for the extreme size of the 1979-flood was thus high precipitation combined with intensive snow melt in higher areas due to high humidity and strong wind. Before noon on August 14th the discharge in Jostedøla river began to rise. After noon the rise accelerated. Cultivated areas were flooded. The water level reached outbuildings and dwelling houses. Roads were damaged and broken. Several people had to abandon their homes in a hurry. In the two hamlets Myklemyr and Fossøy the ground floor of nearly every house was submerged. In some houses the water level even reached the first floor.

In the upper part of the valley the flood culminated just after midnight on August 15th - some hours later further downvalley. The flood showed a very pronounced peak and decreased relatively rapidly after the culmination (see Fig. 3). When the water level had sunk, people in the valley realized the enormous catastrophe the flood had been for them.

- About 100 homes and 30-40 outbuildings were damaged - some few even completely destroyed.
- Great damage occurred over much of the cultivated areas in the valley. Much of the soil was washed away. The crop was wasted
- Most of the bridges in the valley were completely destroyed - among them 6 large bridges that crossed the main river.
- There was considerable damage to the roads. In several places the entire road was totally washed away. The valley was isolated for 10 days.
- All telephone connections was broken.

However, unbelievably enough that no people were killed or hurt.

In the lower part of the valley damages was less. The river Leirdøla is developed for waterpower, and the main reservoir Tunsbergdalsvann was not filled up. When the seriousness of the situation became clear, the hydropower company stopped electricity production and began to store all water from Leirdøla and the glacier Tunsbergdalsbreen in the reservoir. Discharge from 154 km² (18% of the total drainage area) was caught in the
reservoir. At the culmination the discharge into the reservoir amounted to 170 m$^3$ s$^{-1}$. Without this operation - or without the power plant - there would have been enormous damage downstream of the confluence with the river Jostedøla. At Gaupne - the municipality center - the water just reached the top of the levees which were built to protect the town from river floods. Without the manipulation in Leirdøla there would definitively have been severe damages in Gaupne, and many houses would have been taken by water. It is not unlikely that lives could have been lost as well.

**Water discharge**

NVE has installed several water gauges in Jostedalen. The problem, however, was that they were not built for such high water levels. Therefore all records of the flood peak are lost. At two stations, Nigardsvann and Myklemyr (Haukåsgjelet), the course of events has been reconstructed with rather good reliability.

At Nigardsvann, a lake in a tributary river to Jostedøla from west and 35 km upstream the fjord, a limnigraph has been is installed. Nigardsvann drains the outlet glacier Nigardsbreen. The drainage area is 65 km$^2$, of which 74% is glacier covered. Two persons from NVE were stationed there to carry out sediment investigations. When the water gauge broke down, they took manual readings of the water level until the flood culminated. Thus, we know the flood in detail at this location.

In the days preceeding the flood water level in the lake was constant at about 1.80 m, which corresponds to a discharge of approx. 20 m$^3$ s$^{-1}$. Most of the discharge was caused by snow and glacier melting (see Fig 5). At noon on August 14th the water level passed 2.0 m 35 m$^3$ s$^{-1}$). During next 8 hours water level rose a further 85 cm to 2.85 m (89 m$^3$ s$^{-1}$). The culmination took place at midnight. At that time, the water level was 3.08 m and the discharge was 96 m$^3$ s$^{-1}$). During the next 24 hours the water level sunk gradually. At midnight on August 15th the water level had sunk to 2.44 m (48 m$^3$ s$^{-1}$).

The water gauge at Nigardsvann has been running since 1962. The highest water level measured at Nigardsvann apart from during the 1979 flood is 2.75 m, which corresponds to 69 m$^3$ s$^{-1}$.
At Nigardsvann the total discharge during August 14th and 15th is calculated to 9.8 mill. m³, of which, about 50% was caused by glacier melting and 50% by precipitation. That corresponds to a mean precipitation of 70-80 mm in the drainage area during these two days. Most of the precipitation occurred during 12 hours. This agrees fairly well with the precipitation measurements done in the area. The nearest official precipitation station is Fåberg (ca 10 km northeast of Nigardsvann). There it rained 78 mm during 24 hours, which is the highest registered amount during 24 hours since the station opened in 1895. The second highest observed value is 42 mm on August 14th 1898 in connection with the previous highest flood.
Fig. 6  The river Sagarøyelvi transported large amounts of material, even big boulders. It was deposited in the valley near the confluence with Jostedøla where the river gradient decreased and filled up the water bed. Consequently the river broke through the river wall and carried huge amounts of stones and boulders that made great damage on houses and cultivated land. (Photo E.Roland.)

Fig. 7  Myklemyr selfrecording water gauge. At the maximum discharge water level was 1 meter above the floor in the little observation house, or 5-6 m higher than the water level seen on the photo. (Photo S.Krog.)
Myklemyr (Haukåsgjelet) water gauge is located in a narrow gorge in the main river about 15 km from the fjord. The drainage area is 573 km², of which 29% is glacier covered.

In the days before the flood, water level was about 3.0 m, which corresponds to a discharge of about 90 m³ s⁻¹. At noon on the August 14th the water level approached 4.0 m (160 m³ s⁻¹). The next 12 hours the water level rose almost 1 m every third hour and culminated at 8.24 m (766 m³ s⁻¹) at 0200 on August 15th. When the water level passed 7.0 m the instrument was submerged under water and ceased recording. However, the culmination level was clearly visible on the walls in the instrument hut and the instrument started recording again at 0530 when the water level had sunk below 7 m. Thus, the events can be reconstructed with a high degree of reliability. However, there are some uncertainties in discharge values due to a profile change during the flood. Most likely the change occurred near the culmination. Accordingly, the discharge values was calculated from this supposition (Hegge & Krog, 1980). At midnight on August 15th the water level had sunk to 4.50 m (340 m³ s⁻¹). Figure 4 shows that snow melt contributes to about 150 m³ s⁻¹ or about 20% of the discharge at the culmination.

In addition, some of the small tributary rivers from the west, specially Sagarøselvi og Sperleelvi made severe damage. Large amounts of stones and gravel were transported down valley and dammed the river course where the river gradient decreases. The rivers thus broke through their banks and river improvements and damaged cultivated areas - and also some buildings. (See Fig. 4 and Fig. 6.)

At Fossøy (about 25 km from the fjord) an old mark is carved into a vertical rock face, showing the water level during the flood in 1898. In 1979 the water level exceeded that level by more than 1 m.

References

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Hegge, K. & Krog, S., 1981:

Østrem, G., Dale Selvig, K. & Tandberg, K., 1988:
Jostedalen Power Plant
Nils Haakensen, Norwegian Water Resources and Energy Administration

Long time ago it became apparent that Jostedalen and its surrounding glaciers and mountain areas were attractive for water power development. The area has relatively high precipitation, with a specific discharge of 50 - 80 l·km²·s⁻¹.

Planning for the construction of a power plant started in the 1970's by the Norwegian State Power Board. Construction work started in 1984 and the power production started in December 1989. A sketch of the power plant is shown in Fig. 1.

A large dam was built at the outlet of lake Styggevann in the upper part of Jostedalen valley (1157 m a.s.l.). The dam is about 50 m high and contains about 2.5 mill. m³ of rock. A gallery (i.e. a long tunnel) was blasted along the eastern side of Jostedalen. It collects all rivers and most small creeks draining into the river Jostedøla from the east. The lake Styggevann, which is the main reservoir for the power plant, is regulated between 1110 m and 1200 m a.s.l. After the dam was constructed the lakes Styggevann and Austdalsvann formed one lake. At the maximum water level, the reservoir covers an area of 7.8 km² (See fig 2). Approximately 500 mill. m³ water can be stored in the reservoirs, most of it in lake Styggevann. The power plant utilize a head of more than 1100 m (Roland 1988).

Austdalsbreen (not to be confused with Austerdalsbreen) is an outlet glacier from northern Jostedalsbreen that is calving into the reservoir. It is 12 km² and its maximum thickness is about 400 m (see Fig. 3). The calving flux is estimated to be about 200 000 m³ per year (Roland 1986). The periodic rise and fall of the water level in the reservoir has accelerated the recession of the calving front, though a new state of equilibrium has apparently been reached. To prevent drifting icebergs in the reservoir from reaching the dam and causing damage to the dam surface it is covered by a layer of extra large boulders.

In the operation instructions for the power plant it is decided to have a security zone. The reservoir is never allowed to be completely filled up during summer time. The reason for this is flood control, bearing in mind the flood in 1979. Had the power plant been there in 1979, the extent of the flood damages would have been much less.
Figure 1  Principal sketch of the Jostedal Hydropower Project showing the location of Leirdøla, Nigardsbreen, and the Styggevann reservoir.
THE JOSTEDAL PROJECT
Hydropower Reservoirs

Figure 2. Austdalsbreen glacier and the Kupvann and Styggevann-Austdalsvann reservoir

Figure 3. Surface and subglacial topography of Austdalsbreen.
Figure 4. Subglacial topography of the northern part of Jostedalsbreen (from Sætrang and Holmquist 1987).

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Nigardsbreen
Nils Haakensen, Norwegian Water Resources and Energy Administration.

Introduction

Most likely the Jostedalsbreen ice cap totally melted away during the postglacial warm period (8000-6000 BP). Vegetation remnants from this time suggests that the summer temperature was approximately 4° higher than today. Disregarding any possible difference in precipitation conditions this indicate that the ELA (equilibrium line altitude) must have been about 500 m higher than today (Liestøl 1960), and higher than the highest mountains in the area. (See Fig. 2.)

Fig. 1 Map of Jostedalen. Nigardsbreen shown in the box area.
Historical Review

About 5000 BP, climate deteriorated and Jostedalsbreen icecap started to regenerate slowly. There are signs indicating a glacier advance around 3500 BP. About 1500 AD the so called Little Ice Age started. The climate deteriorated substantially and the glaciers increased in size. A lot of minor glaciers in Norway were apparently generated in this period. (Rekstad 1903).

Very little has been written about these events in Norway, mainly because glaciers are located relatively far from populated areas and were generally of little interest to people. However, some documentation exists concerning the last large glacier advance, called “The Little Ice Age,” which culminated about 1750 in Norway. As a result of several years of climatic deterioration, most glaciers advanced considerably, and, in many cases, the outermost moraine at many glaciers today was deposited during this large advance.

The first known written information about Nigardsbreen is found in documents from 1735, when the glacier advanced so far that it started to damage valuable grasslands and even destroyed houses in the valley. It was, for example, mentioned that the glacier in 1735 had advanced so far that it was only a “stone’s throw” from a farm house. In 1743 this farm was completely destroyed by the ice. Until then, it had been the largest farm in the valley, and its name Nigard was later given to the glacier. Other reports mentioned that the glacier advanced almost 3 km between 1710 and 1743. If this information is correct, it indicates that the glacier must have advanced almost 100 m per year during the period 1710 to 1735.

Fig. 2 The relative variation of the ELA in south Norway in postglacial time (from Liestøl 1960).
Fig. 3  Oblique aerial photograph of Nigardsbreen and its moraine system taken by Olav Liestøl in 1951

<table>
<thead>
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<td>- 51 m</td>
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<tr>
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<td>+ 139 m</td>
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* Approximate values only

Tab. 1  Terminus variation of Nigardsbreen during the period 1710 - 1995. Until 1899, when J. Rekstad started systematic front measurements, only scattered measurements existed. After 1899 measurements were done almost every year.
Farmers who lost their fertile land wrote to the king, Fredrik V, and asked for tax exemption because they had lost their grazing fields. The local judge and the minister were sent to the area to inspect the damage. The minister, Matthias Foss, wrote a “Short description from Justedalen 1750,” which was published somewhat later (Foss 1803). He wrote:

In the year 1742, in the middle of August, his Majesty’s representative and the judge went to observe the areas which had been destroyed by the glacier. A measurement was made from the glacier to the nearest house, and this distance was then 200 feet. On the same day next year, 1743, the glacier had not only moved forward these 200 feet, but it also increased in width and had pushed away the houses and tumbled them around. The ice had also plowed up large amounts of soil and gravel and large rocks and crushed these rocks into small pieces which are still visible. The owner of the farm had to leave this house in a great hurry to try to find another place to stay. His farm, named Nigard, was destroyed together with all its fertile land, and the glacier had also approached other areas during the following years. The farm Bjerkehougen lost cultivated land so that only the houses were left, and it is not possible to live there any more. However, we have noted that the ice has retreated since 1748 but only very slowly. In addition to the destroyed arable land the glacier is also harmful because it produces cold winds so that rime on the ground is not unusual during the summer.

In the neighbouring valley, Krundalen, the glacier Tverrbreen made a large advance at the same time, and the officials also visited this place. To register future movements a cairn was erected in front of the ice, and the distance to the front, 19 m, was measured. The distance from the front to a large boulder was also measured. Unfortunately no report of the results of later measurements is available, but this may be the first glacier measurements ever made.

Since this disastrous advance, the glacier has retreated almost continuously, interrupted only by relatively small advances. From 1748 until the present time, the retreat amounts to about 5 km (Table 1). The effect from small advances can be seen easily on the photograph (Fig. 3). The series of small moraine ridges has been dated and described in detail by Fægri (1933), Rogstad (1941) and by Andersen and Sollid (1971). A map showing the various moraines and the time of their formation has been constructed on the basis of all available information. The variation in the front position is graphically represented in Fig. 4.

Continuous records of the glacier fluctuations do not exist. However, single observations
made by visitors to the area have been preserved. C. Smith (1817) studied the glacier in 1812. From his description it is obvious that the glacier was then in a state of retreat which had already taken place for several years. However, he does not give any figures for this retreat.

The next information is given by C. Bohr (1819), who gives an exact figure, e.g. 1726 feet (541 m) for the distance from the outermost moraine to the ice front. He mentioned also that the glacier advanced during the years 1710 to 1742 by about 1/4 old Norwegian mile (almost 3 km).

C.F. Naumann (1824) visited the glacier in 1823 and he gives a less accurate figure of approximately 2000 feet (600-500 m) for the distance from the moraine to the ice front. His description of a very steep front confirms that the glacier was then in a state of advance. A.E. Lindblom (1839) reported that the farmers who lived in the neighbourhood were aware of the
Fig. 5  Map showing the change of Nigardsbreen 1748 - 1974. (From Østrem & al. 1976.)
glacier advance. M.J. Durocher (1847) reports that the distance from the outermost moraine to the glacier was 700 m in 1845. J.D. Forbes (1853), who visited the glacier in 1851, was of the impression that the distance had increased during these 6 years. He made a drawing of the glacier but it is too inaccurate to be used for a dating. J. Rekstad (1902) was told by an old man that the fairly large moraine located about 1600 m from the outermost moraine, was formed during the year of 1873. This is supported by J. Larsen (1874), who stated that the glacier was in a stage of advance during the summer of 1873.

No detailed observations on the glacier's variation were made until Rekstad started his investigations in 1899 (Rekstad 1902, 1904). He applied permanent marks in the bedrock or in very large rocks on both sides of the valley. Distances from these to the glacier front were measured annually in given directions.

However, as soon as the glacier started to retreat across the lake it was possible to use only the northernmost line of measurement, and from 1965 it proved impossible also to use this line. Measuring conditions along the lake were too difficult and, further when the valley

Fig. 6  A vertical profile along the center line of Nigardsbreen showing the decrease of the glacier. In 1748 the ice thickness at the present terminus was more than 300 metres. (From Østrem & al. 1976.)
turned, the direction of the survey line could not be used anymore. After 1973 a system of simple distance measurements was resumed. For the intervening years a photogrametric method was used to determine the position of the glacier front.

The rapid retreat came to an end in 1975. In 1988 the glacier front had its absolute minimum extent (after the little Ice Age). During the past 7 years the glacier has advanced 139 m and is still advancing. A summary of all front measurements (Østrem & al.1976) is given in Table 1. The variation of the glacier size and thickness since the maximum advance is shown in Fig. 5 and Fig. 6.

![Image](image.png)

*Fig. 7* Nigardsbreen painted in 1847 by J.C. Dahl.

Due to the large distance from most glaciers to inhabited areas in Norway only very few historic pictures exist depicting glacier termini. The first known “artistic” representation of Nigardsbreen was made by the famous painter Johannes Flintoe in 1822. The picture shows Nigarsbreen in detail - almost as that of a photograph. Another known painter, J.C. Dahl, portrayed the glacier in 1847. (See Fig. 7.) This painting belongs to the Norwegian National
Later Professor J. D. Forbes from Edinburgh, who visited Norway in 1851, made a good drawing of Nigardsbreen, printed in colours (Forbes 1853) in the book "Norway and its Glaciers". (see Fig. 8.). The very first photography was taken by Selmer and published in "Illustreret Nyhedsblad" (1864).

The first photograph with glaciological intention was taken by J.B. Rekstad in 1899. (See Fig. 9). Later he took several photographs from Nigardsbreen and other glaciers to document glacier retreat or advance. Professor W. Pillewizer took a photograph from Geistdalsnose (1550 m a.s.l.) in 1937 showing the tongue of Nigardsbreen - in the photograph the newly created lake is visible. The first vertical photograph of the tongue of Nigardsbreen was taken by Widerøes Flyveselskap in 1938 in commission of the German expedition. The negatives
Fig. 9  *Nigardsbreen* 1899 photographed by J. Rekstad. The glacier front had then retreated more than 1 km since J.D. Forbes made his drawing in 1851.

Fig. 10  Oblique air photograph showing *Nigardsbreen* in 1947. The lake was still partly ice covered.
were destroyed in Berlin during World War II, but a few positives still exist. On this picture
the lake that appeared for the first time in 1936 is hardly visible. Oblique photos were taken
in 1947 (Fig. 10) and 1954 by Fjellanger-Widerøe A/S. The position of the glacier terminus
was is in the middle of the lake, clearly visible on both pictures.

Three of the last mentioned photographs are reproduced on the back side of the map of

Nigardsbreen is now a prime tourist attraction because, at the end of the 1960's, inhabitants
of the valley constructed a road to the lake Nigardsvann. In 1970, when the lake had been
completely ice free, a boat started to carry tourists across the lake to the glacier terminus.
During the last 30 years tourism has exploded. Today Nigardsbreen is one of the best known
glaciers in Norway. It is visited by thousands of tourists every year. Figure 11 shows
Nigardsbreen as a tourist object.

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Fig. 11  Nigardsbreen today. (Photo Nils Haakensen.)
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Lichenometry used for dating moraine ridges was demonstrated by Beschel (1950), introduced in Norway at Nigardsbreen by Stork (1963), and used later by Andersen and Sollid (1971). A lichen growth curve shows a functional relationship between lichen diameters and the relative ages of moraine ridges. It is valuable to know the age of some ridges to calibrate the curve, and this is possible at Nigardsbreen. A new growth curve (Fig. 1) is established by Erikstad and Sollid (1986) and by means of this curve the moraine ridges can be dated.

**Fig. 1** Lichenometric dating curve (Rhiz. Geog. & alp.)

B is Mountain curve.

C is Nigardsbreen curve.

(From Erikstad & Sollid 1986)
Fig. 2 Scoured surface on the north-east side of Nigardsbreen. White striations are clearly shown. The inner surfaces of crescentic gouges are brown coloured. (From Andersen & Sollid 1971)

Fig. 3 Fluvially polished gneiss surface in the glacier forefield of Nigardsbreen where the glacial striations are almost eradicated. The channel leads down into a pothole. (Sollid 1980.)
**P-Forms.** The term p-forms was introduced by Dahl (1965) to denote "plastically sculptured detail forms on rock surfaces". It should be pointed out that so-called p-forms usually have characteristic laminar stripes imposed on their plastic sculpture.

A number of authors have discussed the origin of such forms, and two points of view are pre-eminent. The one advocates abrasion caused by a glacier and the other fluvial action. Johnsson (1956) thought that a plastic ice mass occurred in subglacial depressions due to pressure melting, and that this ice could move independently of the glacier above. Gjessing (1966, 1967) stated that more or less saturated moraine moving in a viscous manner produced both p-forms and laminar stripes.

Dahl (1965) and Holtedahl (1967) were of the opinion that such forms were caused by fluvial action, and Dahl assumed that the stripes were due to fluvial corrosion. Andersen and Sollid (1971), who described p-forms in front of the retreating Nigardsbreen, showed, however, by means of plaster casts that the same stripes were of glacial origin. The forms were always located in the path of ice-directed drainage, and the rock surface was fluvially polished unless it had been invaded by a later ice stream. Fluvial polishing and glacial grinding in p-forms can alternate at relatively short time intervals, due to advance or retreat of the glacier, or by changing water paths. Seasonal variations in ice velocity also affects the degree of contact with the glacier bed. Glacial reworking of originally fluvially rounded material (Sollid 1964) may be connected with such seasonal alternations.

The rock surface around p-forms has normal glacial striae. Where the ice met obstructions, the glacial striations are deflected due to the forcing of constricted ice around the obstacle.

Plastically sculptured detail forms are observed in bedrock in areas that never were glaciated. An example of this is described from New Zealand (Sollid 1975 a). Plastic p-forms were there sculptured on large boulders by breakers which wash up beach materials with great force. The surface of these features was polished similarly to recent glacifluvial p-forms. P-forms without laminar stripes can thus be formed by non-glacial processes. When a glacier advances over p-forms in bedrock, it is to be expected that ice will be forced into their numerous niches and channels, were it will flow in a laminar manner in the direction of force, thereby producing stripes or striae. As contended by Ebers (1961) p-forms appear to be the
Fig. 4. The roche moutonnée in the foreground has a steep proximal side. The striation is shown. The view is down-valley in Nigard valley. (From Andersen & Sollid 1971)

Fig. 5. Schematic sketch of the proximal part of a roche moutonnée in the Nigards valley. On the steep part of the surface, the striations lie almost at right-angles to the main striation direction of the valley, and are continuously deflected into line with the main striation direction of the valley. (From Andersen & Sollid 1971)
result of two processes: they are excavated fluvially, and their stripes are produced by glacial action in a later phase. Sollid (1975a) emphasizes that the processes are assumed normally to alternate with summer and winter.

**Brown-coloured rock surfaces.** In front of Nigardsbreen brown-coloured rock surfaces occur exclusively in depressions or in the lee of small rock hindrances. They occur beneath the glacier and decrease both in frequency and degree of pigmentation toward the 1750 moraine ridge. The degree of weathering of the rock surface, on the other hand, increases toward the moraine. It appears then, that the colouring disappears with time, and that it is not a weathering product, but a rust-coloured skin.

Studies of a number of cross-sections through this skin confirm this. Rock crystals at the surface show no signs of weathering. The skin, however, is so thin that its chemical composition cannot be distinguished from that of the rock by X-ray analysis. It is assumed to consist of iron oxides stemming from magnetite and pyrite within the rock and is deposited in a very short time with sulphur as the catalyst. Even on proximal surfaces, small brown spots were observed around crystals containing iron.

It is thought that the iron minerals necessary for formation of the skin are derived from the bedrock by glacial erosion, i.e. distal surfaces. Continuous striations are produced when the ice at its thickest and in more direct contact with the valley floor. As the glacier thins, the formation of cavities increases on the lee side of hindrances, where the brown skin is deposited. This is sporadically scoured by striations, which are the very youngest in the area and were possibly formed during the last winter season before the locality was deglaciated. The mode of deposition of the brown skin described here prohibits its direct use as an indicator of the amount of erosion during the last glacier advance.
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Introduction
Nigardsbreen is the largest outlet glacier from Jostedalsbreen. Its surface area is 48 km² (Østreim, Dale Selvig & Tandberg 1988). Nigardsbreen extends from 1950 m a.s.l. down to 350 m a.s.l. Most of its area is part of the large plateau of Jostedalsbreen ice cap (487 km²). Only 6% of the area is below 1200 m a.s.l.

The glacier is quite well known; it is mentioned in written documents from the beginning of the 18th century when the ice tongue started one of the most dramatic glacier advances ever known in Norway. This, as well as its accessibility, may be the reason for much of the scientific activity related to this glacier.

Earlier glacier investigations
Nigardsbreen has been studied by glaciologists, botanists, and others throughout the 19th and 20th centuries. Already in 1812 C. Smith studied Nigardsbreen (Smith 1817). From his description it is obvious that the glacier was then in a state of retreat which had already taken place for several years. C. Bohr (1819) gave an exact figure, 1726 feet (541 m) from the outermost moraine. M.J.Durocher (1847) reported that the distance from the outermost moraine was 700 m in 1845. A. Blytt’s (1869) description from 1864-67 indicate definitely that the glacier then was in a state of retreat. At the turn of the last century J.Rekstad made some scientific studies around Jostedalsbreen, also at Nigardsbreen (Rekstad 1902, 1904). He established permanent survey points in bedrock on both sides of the valley and started annual front measurements in 1899 which have continued nearly unbroken since then. K. Fægri (1933) made some studies of the glacier retreat related to plant invasion in the deglaciated area. In the late 1930-ies the Austrian W. Pillewizer (1952) and the German W. Evers (1939) surveyed Nigardsbreen and other outlet glaciers from Jostedalsbreen.

Andersen & Sollid (1971) made a tedious work to survey and date the moraine system between lake Nigardsvann and the outermost 1748 moraine, using lichometry principally (Fig.1). Several marked moraines are dated, with the 1875 moraine ridge being especially pronounced. It rises 10 - 15 m above the valley floor and starts at a rock knob. It is obviously
the result of a short advance (some few years) that pushed newly deposited and unconsolidated material together and made this big moraine. Another pronounced ridge is the 1930 moraine that is a result of the glacier advance which lasted from 1920-30. This moraine now forms the natural dam of the lake Nigardsvann that appeared for the first time in 1936. Fig. 1 shows the moraine system. Fig. 5. on page 101 shows how the glacier has changed.

Mass balance studies

In 1962 the Norwegian Water Resources and Electricity Board (NVE) started detailed glacio-hydrological measurements for various hydrologic calculations to be used in the planned exploitation of hydroelectric power in the Jostedalen area. When the investigation started a water gauge was also installed in lake Nigardsvann to measure the discharge from the glacier. Since the lake became ice-free in 1967, the total discharge from the glacier is also measured in the glacier stream just below the glacier terminus. Østrem & Karlén (1963) started mass balance studies and discharge measurements at Nigardsbreen and compared the discharge from the highly glacierized basin Nigardsvann with glacier free basins in the same region. The results were used to calculate normal run off values (Østrem 1973). Since then, detailed mass balance studies and water discharge measurements have been carried out continuously.
Fig. 2  Presentation of the mass balance result from Nigardsbreen 1993 (from Haakensen, 1995). The diagram shows how winter, summer, and net balance for 100 m intervals, area net balance, and area distribution. The area net balance curve (Bn) indicate a considerable positive net balance. Only small areas below 1300 m a.s.l. have negative net balance. Equilibrium line altitude - ELA - is at the point where the net balance curve intersects the y-axis (here about 1300 m a.s.l.).

The results from 1962 and 1963 were published in “Norsk geografisk Tidsskrift” (Østrem & Karlén 1963 and Østrem & Liestøl 1964). Since 1964, the results from the investigations has been published in annual reports from NVE in the series “Reports from Hydrologisk avdeling” and from 1984 in “Publications from Hydrologisk avdeling”. The last results available are published by Haakensen (1995). Fig. 2 shows a mass balance diagram as presented in the reports. Fig. 3 shows the mass balance results for the investigation period.
Mass balance - Nigardsbreen 1962-95

Fig. 3  Histogram showing annual winter, summer, and net balance for the entire investigation period (1962 - 95) at Nigardsbreen. Note the excess of positive years in the 1960's and after 1988.

Nigardsbreen 1962 - 93
5-year mean

Fig. 4  5-years running mean for winter, summer, and net balance during the investigation period at Nigardsbreen. Note the marked increase in net balance after 1988.
During the investigation period (1962-95) the mean winter balance measured at Nigardsbreen was 2.25 m, the mean summer balance was 2.00 m, and thus the mean net balance was +0.25 m water equivalents. The cumulative mass surplus since 1962 is 815 mill m$^3$ of water or a total of 17 m water equivalent. The mass balance results are given in Fig. 3. The winter balance varies between 1.52 m in 1977 and 4.05 m in 1989, the summer balance varies between 0.63 m in 1962 and 3.26 m in 1969. The most negative year was -1.31 m in 1969, while 1989 was the most positive year with +3.20 m water equivalent. Fig. 4 shows the 5-years running mean values for the mass balance. It clearly demonstrates how the mean winter balance has increased since 1988. The mean summer balance shows a steadily decreasing tendency from 1980. The net balance shows a greater variation. Until 1988 it undulated between -0.2 m and +0.5 m water equivalents. After 1988 the mean 5-year net balance has been more than 1.0 m of water equivalents.

Fig. 5  ELA as a function of net balance. Note the linear relationship. A similar pattern is found at several other glaciers.
Surface Velocity

Several measurements have been made to determine surface velocity in the ablation area of Nigardsbreen. The first measurements were made by the German expedition 1937 and 1938 which reported daily velocities ranging from 32 cm to 63 cm (Pillewizer 1950).

Fig. 6 Results of surface velocity measurements on the tongue of Nigardsbreen, made in 1937 (dotted line) and in 1951 (solid line). The maximum velocity in each profile is emphasized by an arrow. Numbers indicate daily velocity in cm. From Østrem & al., 1976.)
Olav Liestøl did some measurements in the period 1949 - 1961. He found daily velocities ranging from ca 10 cm at the lower part of the tongue up to 140 cm at the foot of the steep icefall at 1000 m a.s.l. (Østrem, Liestøl & Wold 1976). In Fig. 6 Liestøl’s results are compared with Pillewizer’s.

NVE did comprehensive measurements of surface velocity from 1966 to 1969 (Pytte 1970). The results agree very well with those reported by Liestøl and the Germans. Nielsen (1970) also found that the summer velocity was higher than the winter velocity on the tongue. Measurements on stakes in the firm area (1600 - 1800 m a.s.l.) showed that the velocity was only 0.5 m per year in the highest areas, increasing to 14 m per year in the lower part of the plateau.

Fig. 7 Subglacial topography for Nigardsbreen and the northern part of Jostedalsbreen. (From Sætrang & Wold, 1986.)
Ice Thickness

Ice thickness in the upper part of Nigardsbreen has been surveyed by radio-echo soundings performed by NVE in the 1980ies. A map of the subglacial topography based on these measurements (Fig. 7) shows a highly broken landscape with deep valleys and high mountains (Sætrang & Wold 1986). Ice thickness varies from 20 - 30 m at the lower elevations to more than 500 m under the large flat areas in the upper central part of the accumulation area. In addition, the ice thickness at the tongue (ca 700 m a.s.l.) was determined by hot-point drilling in 1973. Drilling was performed as a part of a survey of the glacier bed, in connection with a planned but never realized subglacial water intake for water power production. The thickness was found to be about 200 m on the deepest part of the lower glacier tongue.

Mapping of Nigardsbreen

The first map of Nigardsbreen was produced by R. Finsterwalder (a member of the German expedition) in 1937-38 by means of terrestrial photogrammetry (Pillewizer 1950). The map covered the glacier tongue up to approximately 1200 m a.s.l. Olav Liestøl constructed a map of the same area in 1951, also by means of terrestrial photogrammetry. This map was never published.

The first modern map of the entire glacier constructed from vertical air photographs was produced in 1964 based upon pictures taken in 1955 and 1964. In 1966 a complete air photo coverage was made for the Norwegian Geographical Survey and a standard topographical map was constructed. The section of this map that comprised Nigardsbreen, was enlarged and formed the base map for the field work until 1974. In 1974 a special air photography was performed to cover the Nigardsbreen drainage basin. A map in the scale of 1:20 000 was constructed and issued in 1975. Finally, a new vertical air photo coverage was made in 1984. Based on these photographs the latest map of Nigardsbreen was constructed and issued in 1988. This map is enclosed with this guide. It summarizes much useful information about Nigardsbreen on the reverse.

The non-glacierized areas on these modern maps were compared with those on the map constructed by Finsterwalder in 1938 and by Liestøl in 1951. The comparison proves that
Glacier surface changes along selected profiles

Fig. 8  Change in surface elevation 1937 - 1984 at four different cross sections on the tongue of Nigardsbreen.
these two old maps have a surprisingly good accuracy. The contour lines coincide very well. A simplified version of these maps are reproduced on the back side of the enclosed map.

The good accuracy of these old maps makes it possible to determine how the glacier surface (below 1200 m a.s.l.) has changed since 1937. Four cross sections of the tongue (ca. 1000 m, 800 m, 600 m, and 400 m a.s.l.) are compared and showing clearly how the glacier surface has lowered since 1937. Above 600 m a.s.l., however, the surface has increased again between 1974 and 1984. This is a consequence of the cumulative positive net balance of the glacier since the mass balance studies were started in 1962. After 1984, the increase in thickness has progressed downglacier, and reached the terminus in 1989 (Fig. 8) involving a glacier advance that has continued until now. Due to the long reaction time (at least 20 years, probably more) the large positive net balance of Nigardsbreen after 1988 definitely will involve continued glacier advance in the nearest future.

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Studies of Water Discharge and Sediment Transport at Nigardsbreen.
Nils Haakensen, Norwegian Water Resources and Energy Administration.

Lake Nigardsvann was born in 1936 in the wake of the retreating Nigardsbreen. By 1967 the lake basin was totally uncovered. The lake is 1.8 km long and up to 400 m wide, and with a maximum depth of 30 m (Fig. 1). Probably the lake was more or less completely sediment filled before the Little Ice Age (Bogen & al. 1989) and was regenerated by advance of Nigardsbreen during the 17th and 18th century.

Water discharge from the lake Nigardsvann has been recorded automatically since 1963. In 1968, after the glacier withdrawal from the lake basin was complete, NVE started sediment transport studies at the inlet and outlet of the lake. The purpose of the investigation was to

Fig. 1  Bathymetric map of lake Nigardsvann surveyed in 1966 when the lake basin was nearly totally uncovered by ice. (From Ekman 1969).
Fig. 2 Water discharge and sediment transport diagram from Nigardsvann 1973. The heavy line shows daily water discharge out of the lake. The open and filled columns indicate the suspended load into and out of the lake, respectively. The difference between these two columns demonstrates the sedimentation in the lake.

collect background data for the engineers who were planning a future water power development. In the period 1968-1981 the field station was manned by assistants during the entire summer and sediment samples were taken five times a day. For these years a complete data set exists for water discharge and sediment transport both at the inlet and the outlet of the lake. Thus it is possible to calculate the rate of sedimentation in the lake. After 1981, samples have been taken by an automatical sampler at the lake inlet only.

These studies have provided a useful record of the variation of sediment transport and concentration in relation to discharge. The sediment concentration usually increases, with increasing discharge, attaining a maximum before discharge maximum (See Fig. 2.). Accordingly, sediment transport increase considerably in flood situations. The mean sediment
concentration is 50-70 mg \cdot \text{l}^{-1}. The highest concentration found any single year range from 360 mg \cdot \text{l}^{-1} to 3000 mg \cdot \text{l}^{-1}. During the 1979 flood the sediment transport during two days (Aug. 14th and 15th) was 9600 tons, a significant fraction of the total annual transport of 18 400 tons. In one single sample taken during the 1979 flood the sediment concentration was 3170 mg \cdot \text{l}^{-1}. During the period 1968-94 the total transport of suspended material from Nigardsbreen was 295 000 metric tons (approx. 150 000 m³) or an average of 10 900 tons annually. Until 1981 the transport of suspended material out of the lake was also measured to calculate the sedimentation rate. The percentage of the sediment that was deposited in the lake proved to be very stable, varying between 75 and 80% of the input. The percentage typically increases with increasing discharge.

In 1969 an attempt was made to measure the total amount of coarse material carried as bottom load. A large vertical fence was installed across the entire river. The fence consisted of a heavy steel net with a mesh size of 2 cm. The net was kept in position by 25 steel rails, 3 m long, drilled into the bedrock and supported by five 1-inch prestretched steel cables (Østrem 1975). During the period from May 24th to June 19th, 398 metric tons of material coarser than 2 cm was trapped in the net. During the same period 1276 tons of suspended material was measured in the river (Østrem & al. 1970). During a flash flood at the end of June the installation was totally destroyed.

Most of the water discharge and almost all sediment transport occurs during the summer months. The winter discharge is normally in the order of 0.1-0.4 m³ \cdot s^{-1} which is negligible compared to average summer discharge, about 20 m³ \cdot s^{-1}. During the flood of 1979 a maximum discharge of 96 m³ \cdot s^{-1} was observed. This corresponds to nearly 10 mill m³ pr. day.

The discharge show great variations from one year to the next. The highest observed summer discharge was 244 mill. m³ in 1969. That year the glacier had a substantial mass deficit amounting to 63 mill. m³. The lowest discharge occurred in 1987 with 124 mill. m³. That year the glacier had a large positive net balance and 75 mill. m³ water was stored on the glacier as snow and ice.

As stated the lake became completely uncovered in 1967. Immediately, a delta started to grow at the lake inlet. In 1968 the delta area was carefully levelled along profiles between a
number of fixed points established around the delta area. Since then, repeated levelling and surveying has been carried out every year along identical profiles. Thus, the amount of coarse material deposited at the delta can be calculated. Accordingly, we have a detailed knowledge of the delta growth since the delta was born in 1967.

Fig. 3 shows the delta growth during the investigation period 1968-94. Fig. 4 shows a bathymetric map from four different years showing clearly how the delta has changed since 1968. The delta area, which is defined as the area which becomes dry at low water level, has increased from 23% of the inner lake basin in 1968 to 68% in 1994, whereas areas deeper than 2 m have decreased from 43% to 11%. In the same period the delta growth amounts to 168 000 m$^3$ or 6200 m$^3$ pr. year.

Fig. 3. Delta growth 1968 - 1994 (from Haakensen & al.). As the process continues, the area with maximal sedimentation will move down the lake.
Today the water volume in the inner basin is approx 55,000 m³. If the delta growth continues at the same rate in the future, the inner basin will be filled up in about 10 years and the delta growth will enter the outer basin (at the map's right margin).

In 1968, 1978, 1980, 1983 and 1991, a large number of sediment cores were taken from the lake bottom. In total cores were taken from 38 different sites in the lake.

In 1968 cores were taken from 15 different locations. During this winter expedition a large amount of brick powder was introduced at a number of the sampling sites. All this was done from the solid ice cover and thus the locations could be surveyed with good accuracy. The brick powder was found by later sediment sampling, and can thus be used as a time marker.
Fig. 5. Sediment cores from Nigardsvann.
a) Sample obtained by piston core in 1968.
b) Sediment core taken in 1978. The glacier front position was here in 1943.
The brick powder introduced in 1968 is visible in the core.
The brick powder layers indicated the 1968 horizon and was used as time markers. Two different cores are shown in Fig. 5. The cores clearly show annual varves and by counting these and measuring the varve thicknesses the sedimentation rate can be calculated. In the outer part of the lake there are more visible varves than in the inner part. The reason is that the lake became uncovered by the glacier between 1936 and 1967. The number of varves in 11 different samples corresponds well with the time period since the corresponding part of the lake basin was first uncovered by the retreating glacier.

Comparison of actual sedimentation volumes from core studies agree well with measured suspended load concentration into and out of the lake. For the 11 year period 1968-79 the total transport into the lake Nigardsvann was about 115,000 tons of which 90,000 tons were deposited on the lake bottom.

During the entire observation period (1968-94) the transport of suspended material from Nigardsbreen was 142,000 tons and the delta growth 168,000 tons. The total transport was therefore 310,000 m$^3$ in 27 years. Dividing by the area of Nigardsbreen gives an estimated erosion rate of 6.5 mm or 0.24 mm·a$^{-1}$. 
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