GLACIOLOGY, HYDROLOGY AND GLACIAL GEOLOGY AROUND JOSTEDALSBREEN

FIELD GUIDE TO EXCURSION
10 SEPTEMBER - 13 SEPTEMBER 1988
ORGANIZED IN CONJUNCTION WITH
SYMPOSIUM ON SNOW AND GLACIER RESEARCH
RELATING TO HUMAN LIVING CONDITIONS
LOM, NORWAY, 1988

NORSK-POLARINSTITUTT

OSLO, NORWAY, 1988

NVE
NORWEGIAN
WATER RESOURCES AND
ENERGY ADMINISTRATION

GLACIOLOGY, HYDROLOGY AND GLACIAL GEOLOGY AROUND JOSTEDALSBREEN

By

I. AARSETH, J. BOGEN, H. HOLTEDAHL,
O. KLAKEGG, A. NESJE, N. RYE,
O. ORHEIM and E. ROLAND

FIELD GUIDE TO EXCURSION
10 SEPTEMBER - 13 SEPTEMBER 1988
ORGANIZED IN CONJUNCTION WITH
SYMPOSIUM ON SNOW AND GLACIER RESEARCH
RELATING TO HUMAN LIVING CONDITIONS
LOM, NORWAY, 1988

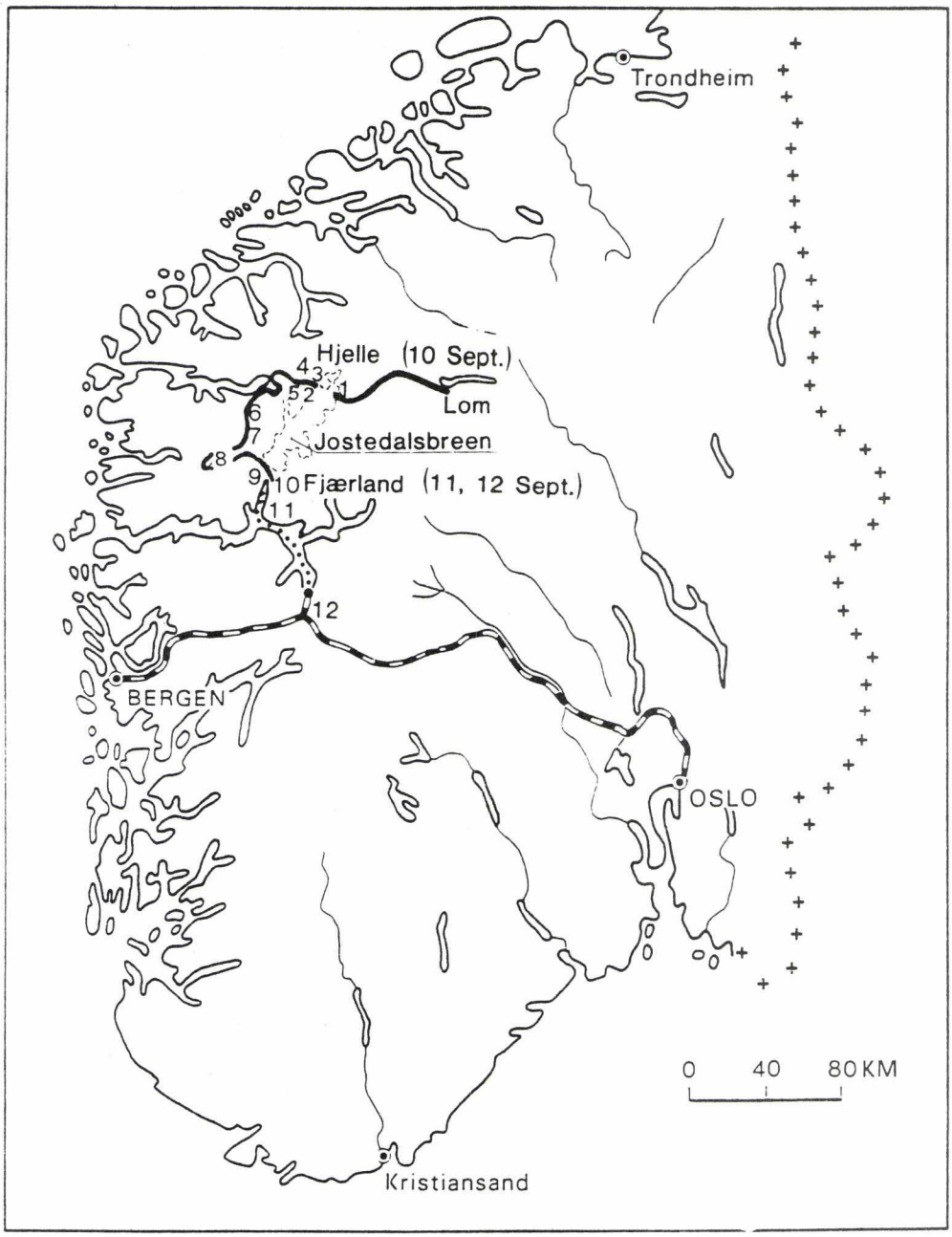
EDITED BY O. ORHEIM

NORSK POLARINSTITUTT

OSLO, NORWAY, 1988

NVE
NORWEGIAN
WATER RESOURCES AND
ENERGY ADMINISTRATION

CONTENTS



Route of the four-day excursion, with numbered localities referred to in the text.

PREFACE

This field guide is made for a four-day excursion taking place immediately after the Symposium on Snow and Glacier Research relating to Human Living Conditions, Lom, Norway, 5-9 September 1988. This Symposium was organized by the International Glaciological Society, Cambridge, England, in cooperation with Norwegian Geotechnical Institute, Norwegian Water Resources and Energy Administration (NVE) and Norwegian Polar Research Institute (NP). Organization of the excursion has involved work of many persons. OLAV ORHEIM (NP) has been responsible for the practical arrangements. The following have responsibility for the scientific content of the various sections of the excursion: I. AARSETH, H. HOLTEDAHL, A. NESJE and N. RYE, Department of Geology, Univ. of Bergen; J. BOGEN, (NVE); O. KLAKEGG, Norwegian Geological Survey; O. ORHEIM, (NP); and E. ROLAND, Norwegian State Power Board (Statkraft).

A decision to visit Brigsdalsbreen was made after this guide was ready for printing. This additional locality, coming between loc. 5 and loc. 6, is described at the end of the guide.

The individual sections of the guide for which the authors are responsible are listed on the Page of Contents. The guide has been edited by OLAV ORHEIM, and EVY MARIANNE VOLLMO, and NILS HAAKENSEN, NVE, did the technical editing. Statkraft, NVE and NP financed the printing of the guide.

Several separate publications are enclosed with this guide:
1) (concern loc. 5): NGI Publ. No 79, which contains the following two articles: Bjerrum, L. and Jørstad, S. 1968: Stability of rock slopes in Norway, 11 p.; and Jørstad, S., 1968: Waves generated by landslides in Norwegian fjords and lakes, 20 p.;
2) (concern loc. 10): Orheim, O., 1970: Glaciological investigations of Store Supphellebre, West-Norway. Norsk Polarinstitutt Skrifter 151, 48 p.
3) Østrem, G., Dale Selvig, K. and Tandberg, K. 1988: NVE Atlas of Glaciers in South Norway, Revised Edition, pp B65-B80.

Several maps are also enclosed, including:
1) 1988 map of Jostedalsbreen;
2) Aa, A.R. 1988: Brigsdalsbreen, 1318 II, kvartærgologisk kart (Quaternary map) - N 1: 50000. Norges Geol. Unders.
3) road maps.

AUSTDALSBREEN AND THE STYGGEVATN HYDROPOWER RESERVOIR

Introduction

Austdalsbreen (loc. 1) is an outlet glacier facing southeastwards of the northern part of the icecap Jostedalsbreen. Today a minor part of the glacier front is calving in the lake Austdalsvatn (1157 m a.s.l.) at shallow depths (max. 10 m). The glacier covers an area of 11.3 km^2 and has a length of 5.6 km along the central flow line from the head of the glacier to its terminus. The equilibrium line elevation is 1460 m.

Winter precipitation comes mainly with low pressures from southwest and west in October - April. Mass balance measurements were carried out in the years 1967-72 on Vetledalsbreen (4.2 km^2), an adjacent glacier facing northeastwards, and on Nigardsbreen (48 km^2 facing southeast) since 1962. Results are reported in annual reports from Glacier Division, Norwegian Water Resources and Energy Administration e.g. Roland and Haakensen (1985). From these data the mean annual mass exchange on Austdalsbreen is calculated to 2.0 m water equivalent. Annual summer ablation on the glacier tongue amounts to 3-5 m. Maps based upon airphotos from 1966 and 1984 (lower part only) show that the calving front receded about 100 m during that period. Mass deficit in lower parts of the ablation area corresponds to a lowering of the surface of 20 m during these 18 years.

The Styggevatn reservoir

The two lakes Austdalsvatnet (1157 m a.s.l.) and Styggevatnet (1156 m a.s.l.) will be dammed up to 1200 m a.s.l. by means of a rockfill dam, to become a reservoir for a power plant. At the maximum water level the reservoir will cover an area of 7.8 km^2 and have a length of 6 km (Fig. 1). Simulations based upon data from 1930 to 1960 show that the future mean water level will be 27 m above the "old" level.

JOSTEDALEN

HYDROPOWER PROJECT

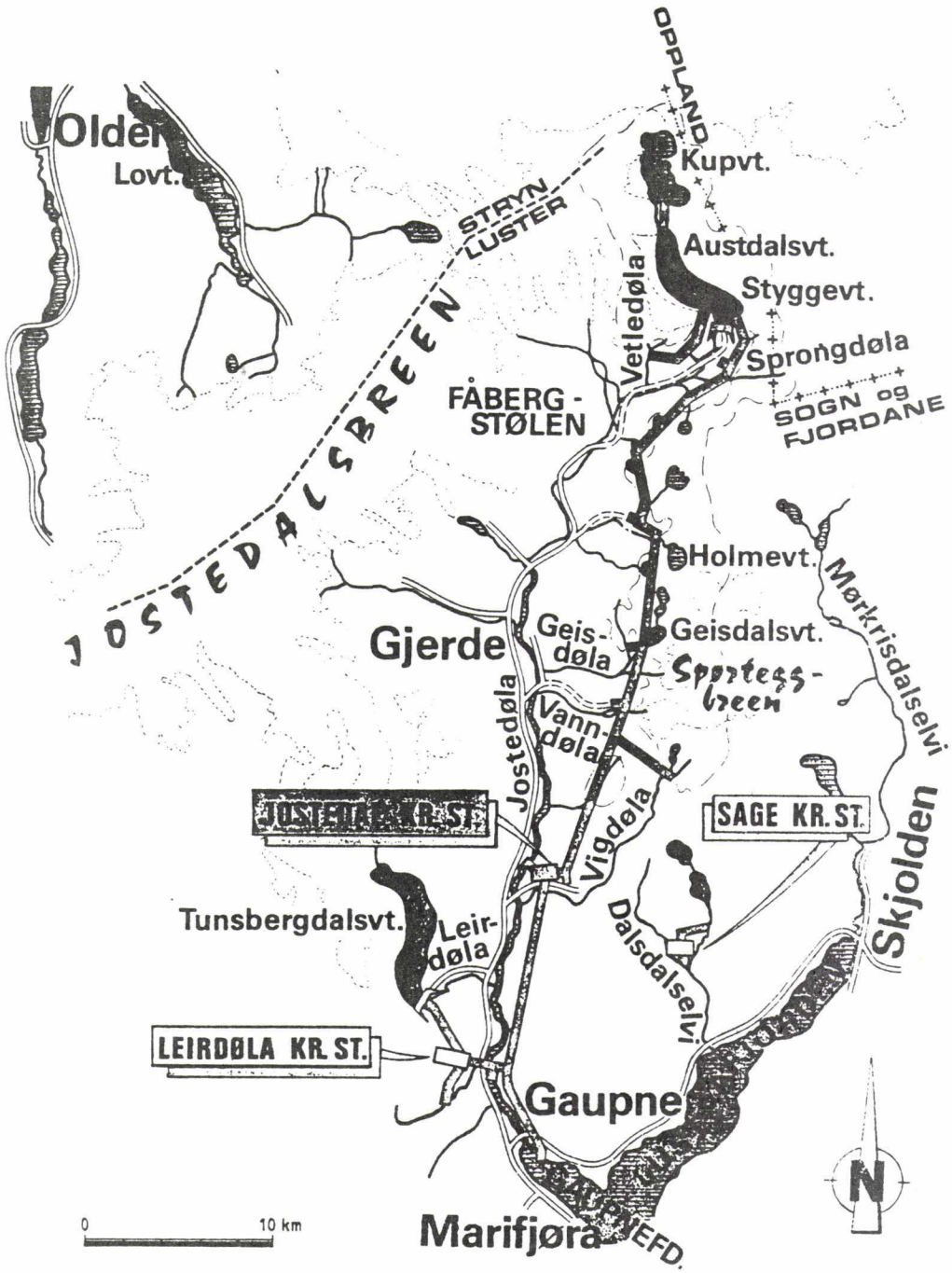


Fig. 1. The Jostedal Hydropower Project and the locations of the Styggevatn reservoir in the upper part of the Jostedalen valley.

Fig. 2 shows the planned dam height and the water level during the filling of the reservoir in 1988 and 1989. Hitherto in 1988 the observed water level has been higher than the estimated one. The daily mean rise from June 16 to August 22 amounted to 0.27 m.

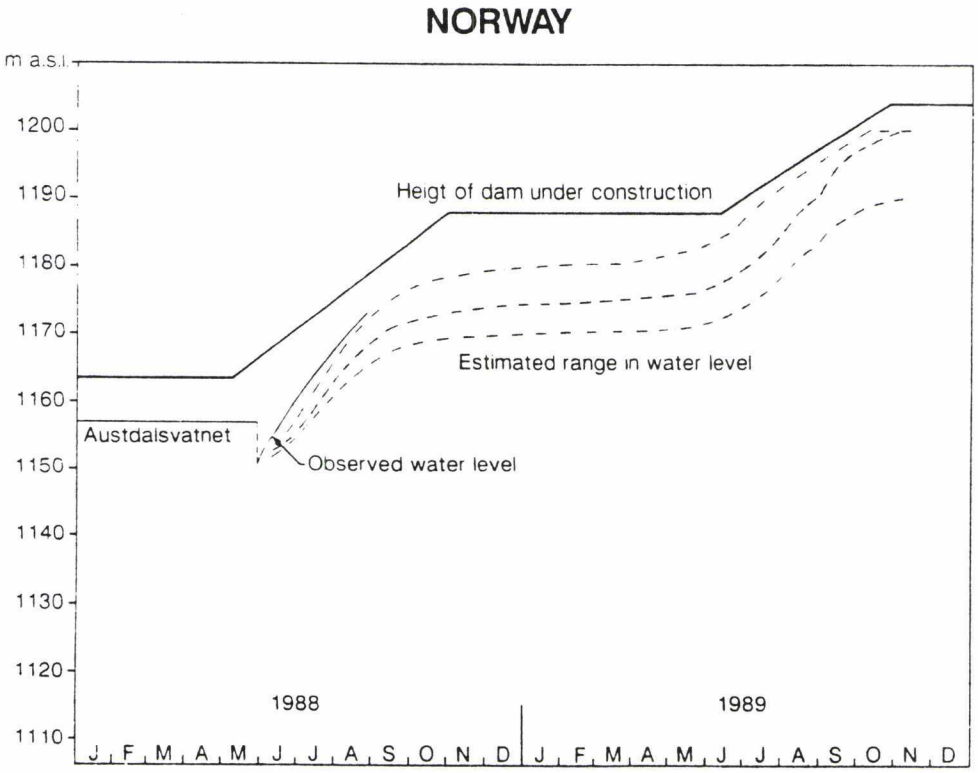


Fig. 2. Observed and estimated (25th, 50th and 75th percentiles) rise in water levels in the reservoir in 1988 and 1989.

Observation programme

A comprehensive glacier observation programme has been initiated in cooperation with several scientists, including Röthlisberger (1987), to predict any dramatic glacier events and changes in the reservoir volume. Also, the State Power Board wish to increase experience and knowledge about glacier dynamics which will be used in future hydropower projects.

Subglacial topography

Glacier thickness of the northern part of Jostedalsbreen was mapped with radioecho-sounding (RES) in 1986 (Sætrang and Holmquist, 1987). The subglacial map shows a deep valley underneath Austdalsbreen. Two thresholds have both saddle points close to the future maximum water level of the Styggevatn reservoir (Fig. 3). Discrepancies between radioecho-sounded bottom topography and bottom in boreholes drilled with a hot water jet initiated a detailed mapping on the outer threshold with RES. Sætrang (1988) stated that the elevation of saddle point was 1215 ±3 m a.s.l. in accordance with the bottom of the boreholes.

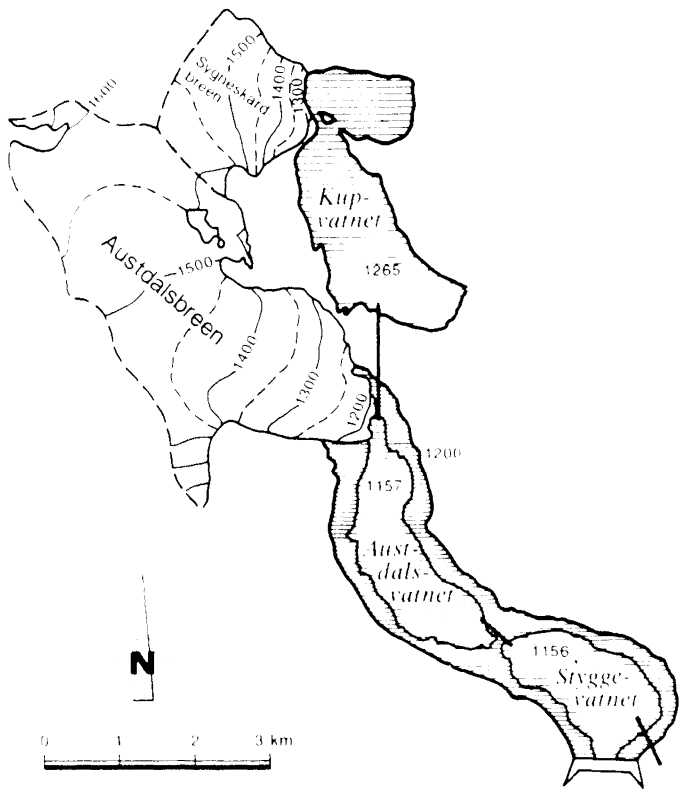


Fig. 3a.Sketch of the reservoir and the calving glaciers.

Subglacial water pressure

Water pressure sensors were installed at the bottom of eight boreholes located within four surface movement grids in October 1987. Hourly observations are registered on dataloggers. Last winter the measured pressure increased to above the ice overburden pressure on the threshold 1.6 km from the glacier front indicating that the borehole was closed off. This high pressure lasted one month from mid-December and reoccurred from the beginning of July, and may reflect local dynamic conditions along the glacier bed. Large daily variations in water pressure occurred from early July at the locations further up- and downglacier, indicating good connections with meltwater channels and boreholes. The location near the front shows small fluctuations in water pressure indicating a "ground water table" in the glacier, see Fig. 4.

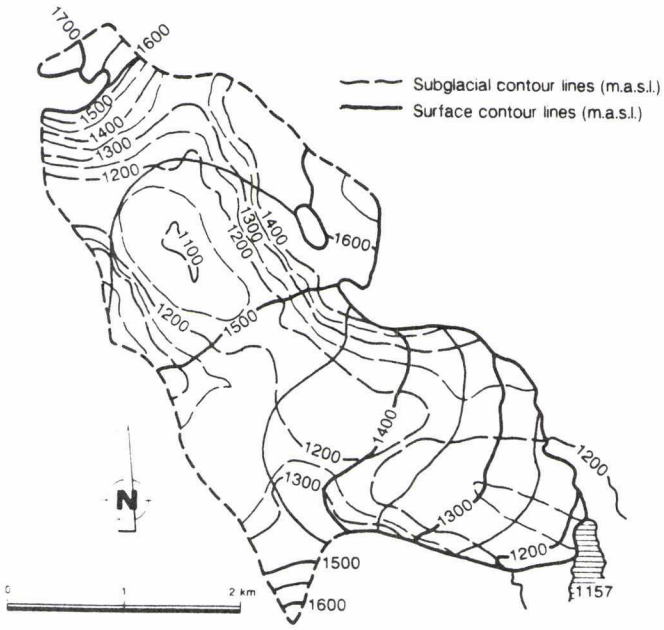


Fig. 3b.Surface and subglacial topography of Austdalsbreen.

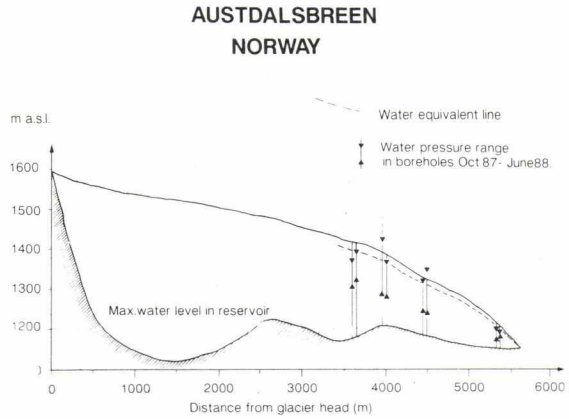


Fig. 4a.Long profiles of surface and topography along the central flow line. The positions of the boreholes and the range in water level pressure are shown.

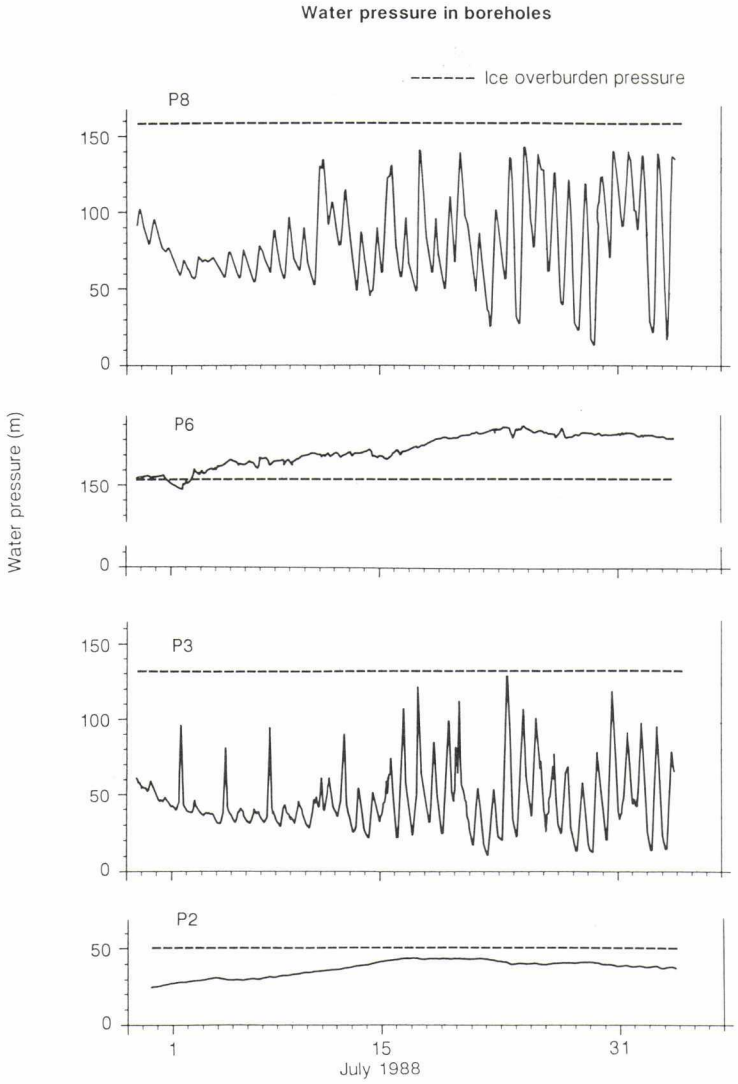


Fig. 4b.Fluctuations in water pressure in boreholes in July 1988

Velocity

Measurements of flow rates to be used in dynamic models started in 1986. Velocity and deformation is observed in cross profiles and grids along the central flow line on the lower parts of the glacier Calculated velocity distributions before the rise in water level are shown in Fig. 5. Horizontal velocity at the calving front amounts to 25 m per year.

Increased water levels in the reservoir will change the physical conditions along the glacier bed. An analogous situation existed in 1957 at Austerdalsisen which calved into a former glacier dammed lake in the Svartisen area in northern Norway. The surface velocity followed the seasonal variation in water level with a short time lag during the drainage of the lake (Liestøl, pers. comm.). We may therefore expect that changing water pressure at the glacier bed of Austdalsbreen will affect the velocity on the lower parts of the tongue. On the average, the velocity will increase.

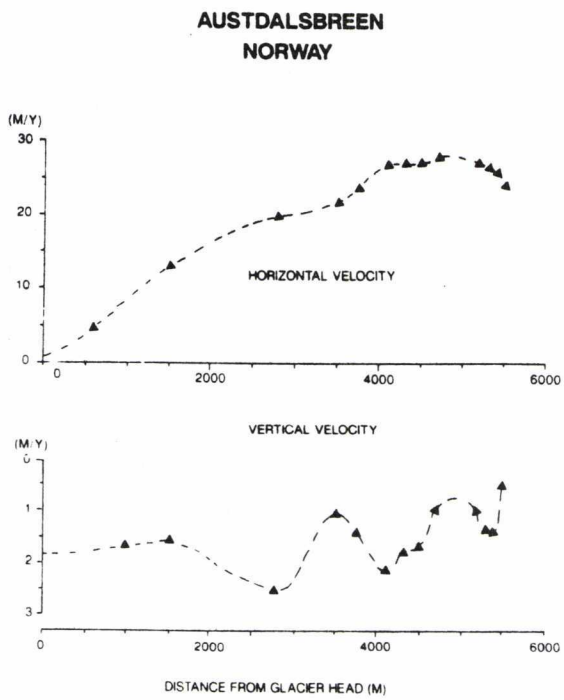


Fig. 5. Velocity distributions along the central flow line.

Calving

The calving flux depends on the physical conditions in the proglacial lake. Among these, the water depth is the most important parameter in freshwater lakes. The ice flow into the Austdalsvatnet is calculated to about 150 000 m^3 per year. The retreat of the calving front in the period 1966-84 amounted to a yearly mass loss of appr. 50 000 m^3. Thus the total calving flux was estimated to about 200 000 m^3 per year, (Roland, 1986). This corresponds to a water layer of 0.02 m distributed over the total glacier area. Ablation due to calving was therefore of minor importance in the mass balance of Austdalsbreen before the lake was dammed.

The rising water level in the reservoir will submerge the lower parts of the glacier tongue and cause increased fracture. A calving bay may develop along the central flow line, and the width of the calving front increase from 300 to 900 m. The calving is expected to increase by an order of magnitude.

Glacier advance or retreat?

The front position of calving glaciers is determined by the ice flow to the terminus and the calving flux. The expected increased calving rate imply that the velocity at the front will have to increase dramatically to cause a glacier advance. Therefore the glacier front is assumed to retreat.

A glacier retreat will increase the reservoir and thus benefit the power production. Experience from other reservoirs with calving glaciers in Norway show that the calving flux exceed the ice flow at the front. Thus the glaciers have receded. Also a climatic warming has caused retreat.

Calving waves

Measurements of calving intensity and resulting water waves in Austdalsvatn were carried out during September - October 1987. A waverider buoy was placed 200 m from the glacier front. The maximum

observed height of the calving waves amounted to 1.3 m. The wave period varied from 2 to 11 seconds. (Steen and Mathiesen, 1987). The most reasonable future maximum wave heights are calculated by Schieldrop (1986) to about 3 m. Such a wave will not cause damage on the rockfill dam, but will obviously affect the near shores of the inner part of the reservoir to a larger degree than today. There are tourist routes crossing this area and the danger will clearly increase.

Icebergs in the reservoir

Drifting icebergs in the reservoir may reach and cause damages on the rockfill dam. This will possibly be prevented by installing a boom or hawser across the reservoir to stop the icebergs 2 km away from the dam.

Sediment transport

Cores from the bottom of Austdalsvatn indicate a production of 1200 tons sediment per year from Austdalsbreen. The mean sedimentation rate, measured 250 m from the glacier front at a depth of 14 m, amounts to 0.6 mm per year in the period 1780-1915, thereafter the average rate is 3.1 mm per year (Bogen and Olsen, 1986). These low sedimentation rates show that Austdalsbreen probably is a "hard bed" glacier. Subglacial depressions may, however, contain postglacial sediments.

VALLEY SANDURS AND GLACIERS IN ERDALEN

The valley of Erdalen (loc. 3) in Stryn represents the upper part of a glacially shaped valley system composed of a number of glacial steps and glacially overdeepened basins (Fig. 6).

Some of the basins are occupied by lakes, but several are filled up by sediments derived from the retreat of Pleistocene glaciers and from sediments supplied from modern glaciers during the Holocene. The development of sandurs, riverplains and deltas reflects the long term variations in sediment supply from the glaciers Vetledalsbreen and Erdalsbreen. These glaciers are outlet glaciers from the Jostedalsbreen ice cap and are situated in the upper part of the valley. In the year 1970, the sediment transport in the

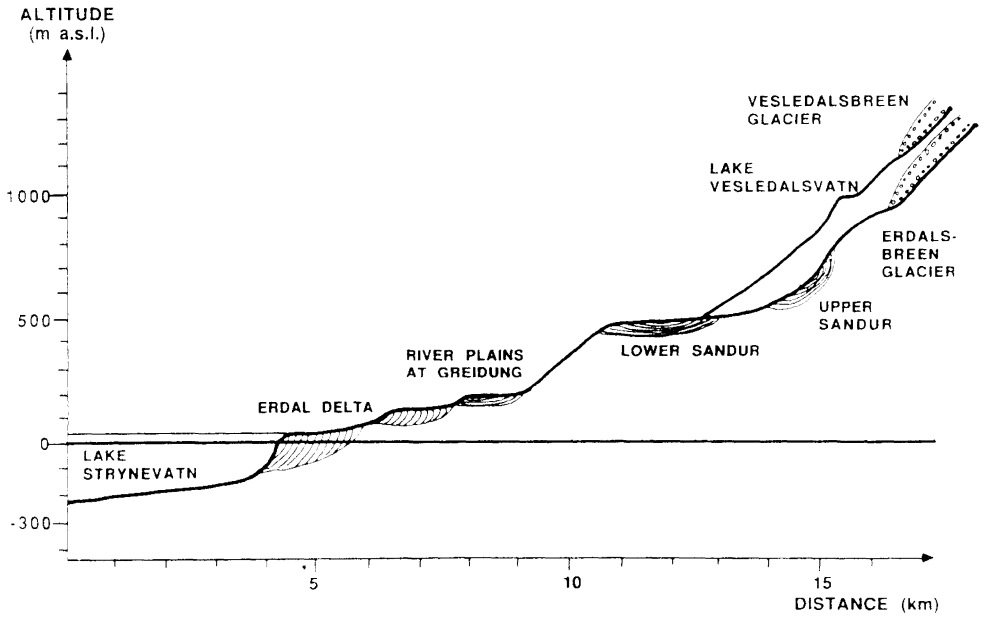


Fig. 6. Longitudinal profile of the Erdalsbreen river system.

meltwater river from Erdalsbreen was measured as 33500 tonnes, of which 19250 tonnes were suspended sediments. In the same year Vetledalsbreen supplied a total load of 670 tonnes, most of which accumulated in the lake Vetledalsvatn (Fig. 7).

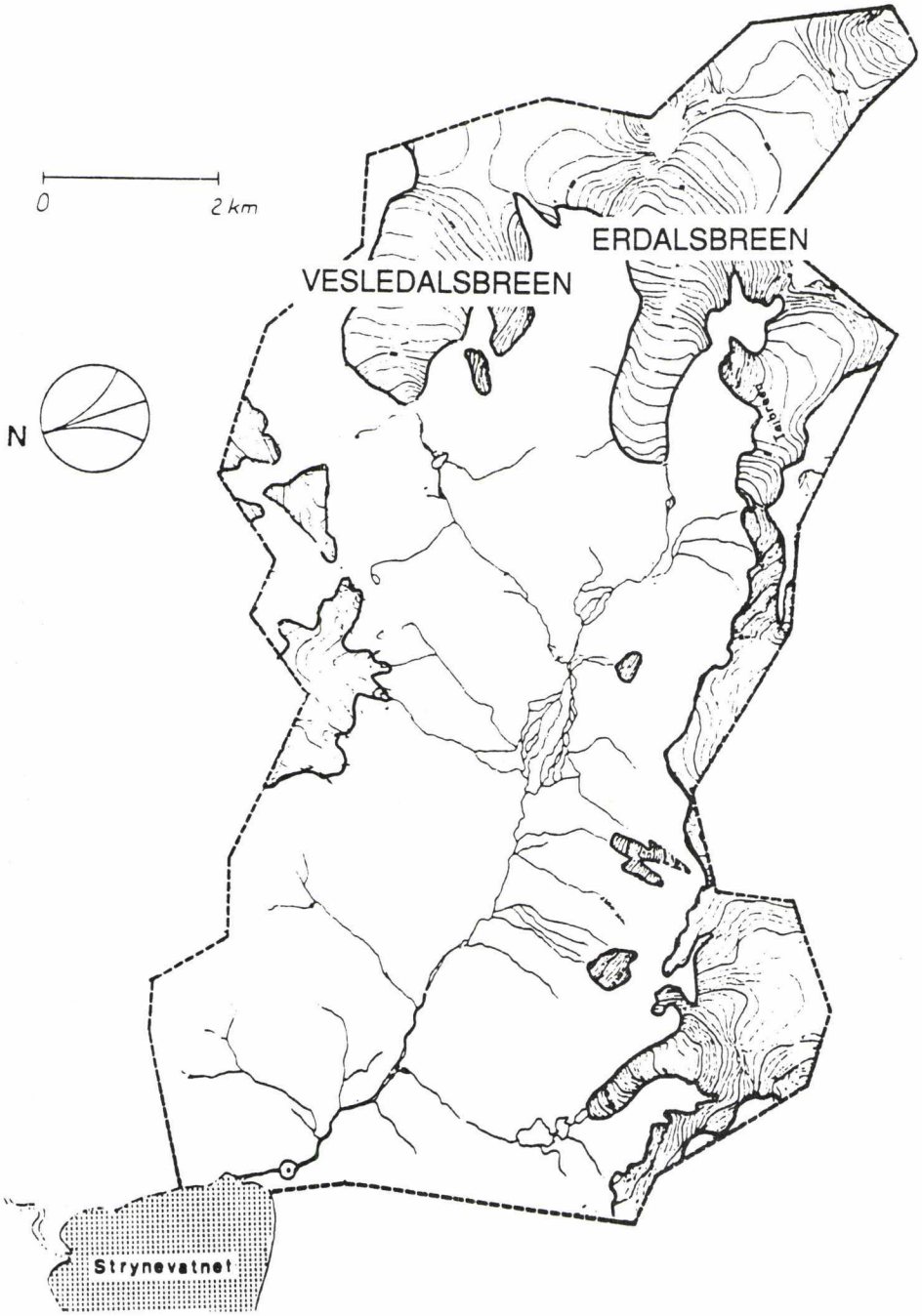


Fig. 7. Location of the Erdalsbre, Vetledalsbreen and the valley sandurs.

Neoglacial moraines dating from around 1750 can be found in front of Vetledalsvatn and at Ulvestigen about 1 km from Erdalsbreen. Two sandurs have been formed (Figs. 7 and 8). The lower sandur is about 2 km long and covers an area of 0,8 km^2, at an altitude of 460-495 m a.s.l. The upper sandur is a small outwash fan, ≈ 1 km long with an area of only 0.2 km^2. The channels are dominated by gravel, cobble and boulder sizes. Raubakken (1985), has investigated the grain size variations of the sandurs (Fig. 9).

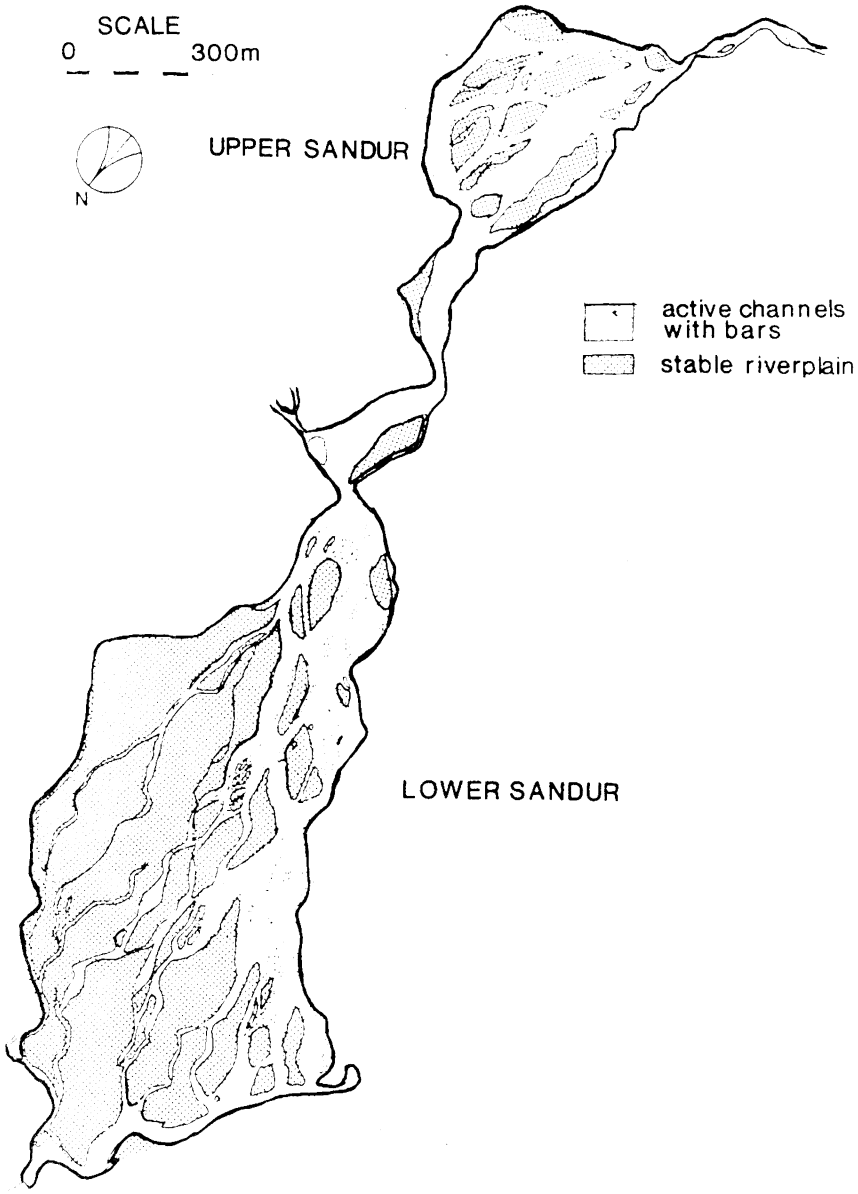


Fig. 8. Map of the upper and lower sandurs.

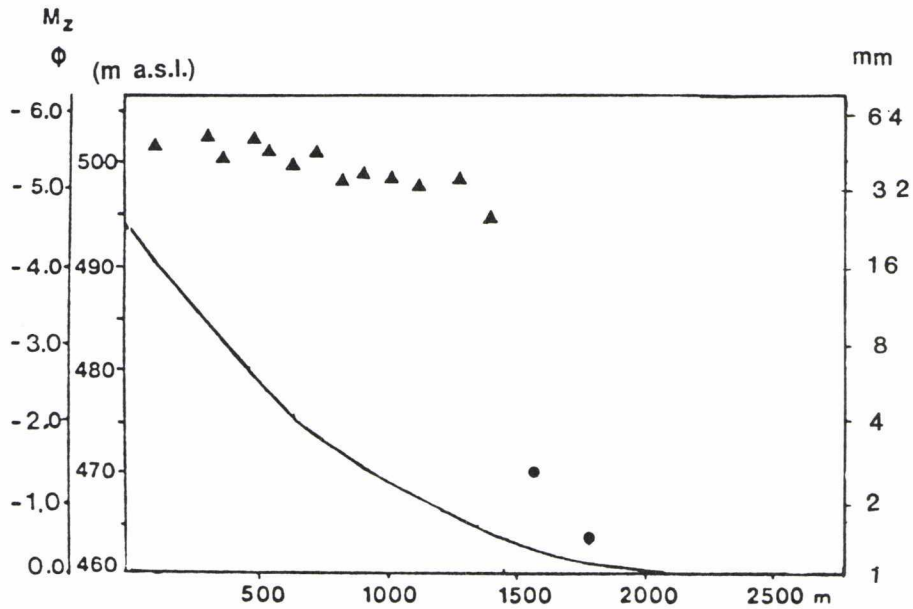


Fig. 9. Downstream variation in grain size distribution of the lower sandur (B). Triangles: Wolman analyses. Dots: Sewving analyses. Solid line: Slope.

Sandurs are braided river systems developed downstream from glaciers. The channels in the sandurs of Erdalen are however confined by valley walls. These type of sandurs are called valley sandurs by Krigstrøm (1962). River channels on sandurs are frequently subject to channel splitting due to heavy aggradation of sediments supplied from the glaciers. This type of activity seem to decrease on the lower sandur in Erdalen, and parts of the system develop into a stable riverplain. The general retreat of the glaciers is most probably the cause of this development. As glaciers recede, a large part of the coarser fractions of the load is deposited upstream in proglacial positions, and the amount of sediments supplied to the sandur areas is reduced. The front variations of the glaciers in Erdalen have not been investigated in detail, but the pattern is most probably similar to that of the measured outlet glaciers from Jostedalsbre and Folgefonni. A general retreat has been experienced from the advance in the 18th and 19th century. During the 20th century the glaciers have generally been retreating, but several glaciers advanced during the

periods 1900-1911, 1923-1933 and 1974-1986 (Fig. 10).

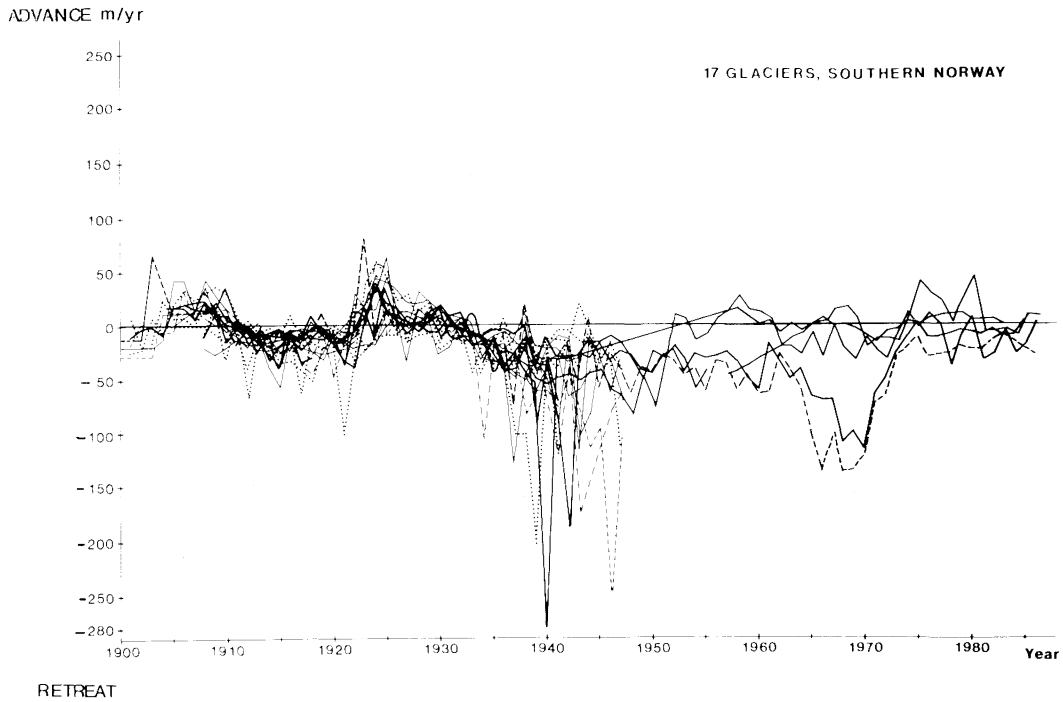


Fig. 10. Front variations of 17 glaciers in south Norway. From Bogen, Wold and Østrem (1988).

In the Alps aggradation and associated densification of the network of braided channels have been observed in front of advancing glaciers (Maizels, 1979).

We believe therefore that the advance of the glaciers in the Jostedalsbre area around 1750 caused similar effects. Channel changes, involving the elevation of river beds and bank erosion, normally lead to increased flooding and damage to farm land. Grove and Battagel (1983) gave a frequency diagram of tax reductions allowed for farmers in the Stryn area due to inundations caused by floods, avalanches, rockfalls, landslides and glacier ice avalanches in the 17th, 18th and 19th century (Fig. 11).

The present proglacial lake that formed in front of Erdalsbreen during the 1980's has reduced the sediment supply to the lower part of the Erdalen river. However, flood protection works would have to

be made to protect houses and farm land if the glaciers were to readvance considerably.

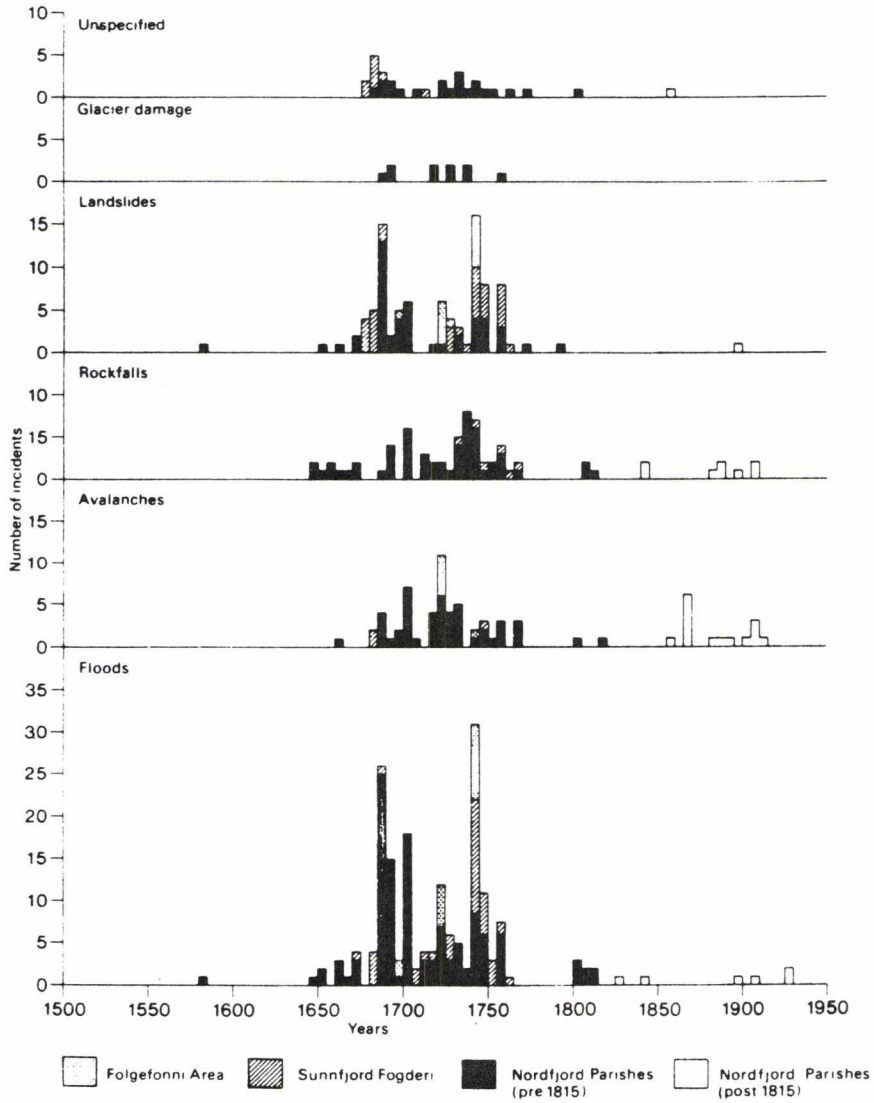


Fig. 11. Frequency diagram of tax reductions due to glacier inundations on farm land on the western side of the Jostedalsbreen ice cap.

A coarse-grained delta is prograding into Strynevatn where the river enter the lake. Strynevatn is more than 200 m deep in this area. A comparison of similar deltas investigated by Bogen (1983) suggests that the irregular form of the delta was caused by Holocene subaqueous slides that deformed the delta front. Raubakken (1985) investigated the sedimentation in Strynevatn with sediment traps and bottom cores in the bottomset beds of the delta (Fig. 12). The deposition rate varies from 800 g/m^2 yr near the river mouth to about 25 g/m^2 yr at 200 m depth 1 km downstream. Mean grain size decrease from 1.0 mm to 0.016 mm within the same distance.

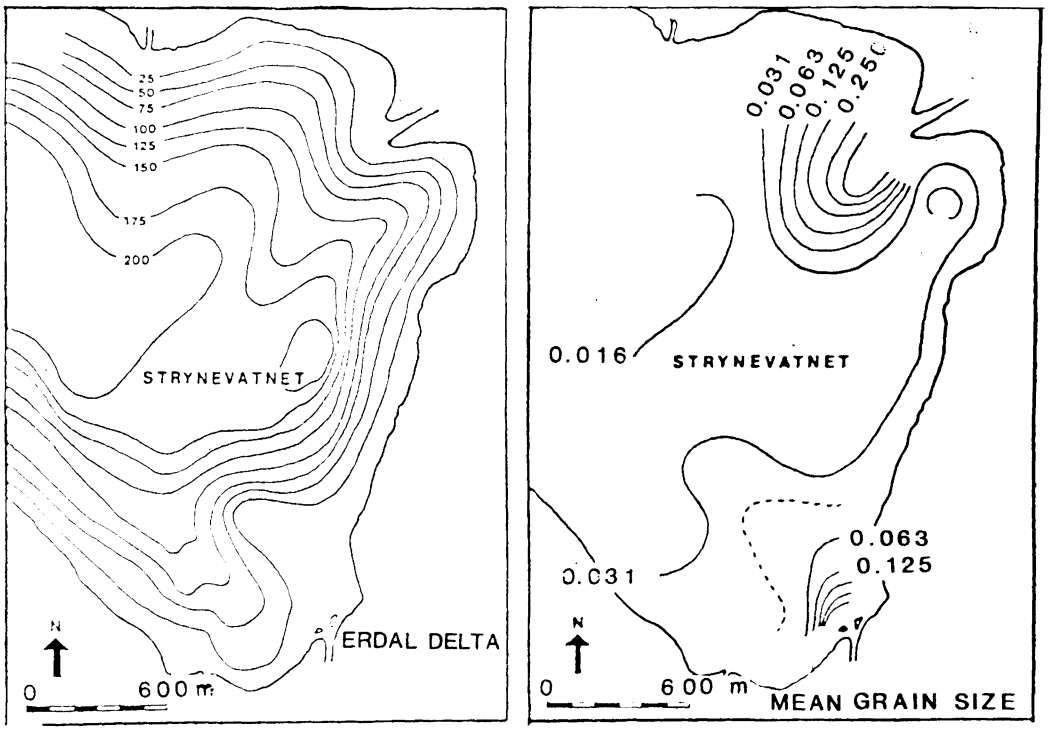


Fig. 12 Bathymetric map of the inner part of Lake Strynevatnet, showing the delta of River Erdalselv. Contour lines in m. Grain size variations (in mm) of the bottomset beds of the lake.

GLACIATION AND DEGLACIATION OF INNER NORDFJORD

Topography and bedrock

In the Nordfjord area (Fig. 13) a high plateau is incised by fjords and deep valleys. The mountain tops are relatively flat or undulating, commonly bordered by steep cliffs. The highest mountain plateaus around Nordfjord descend from about 1800 m elevation at the Jostedal Plateau to 400-500 m a.s.l. at the coast. The bedrock is dominated by gneiss with some massive granite, amphibolite, quartzite, schist, eclogite and dunite.

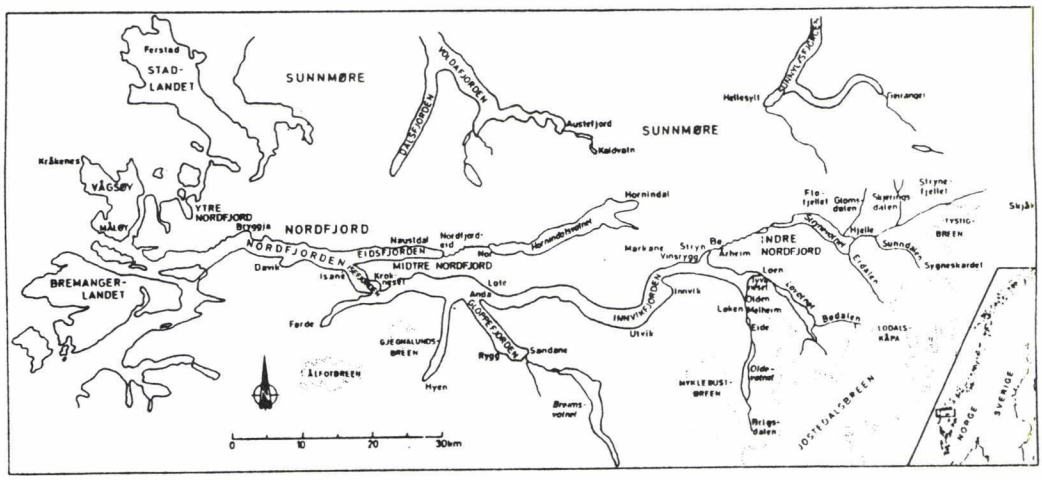


Fig. 13. Location map of Nordfjord. (Rye et al. 1987.)

The last glacial maximum

The highest mountain areas around inner Nordfjord are covered by in situ block fields, and some mountains have alpine morphology (Fig. 14). The block fields commonly have sharp lower boundaries which indicate an upper limit of regional glaciation.

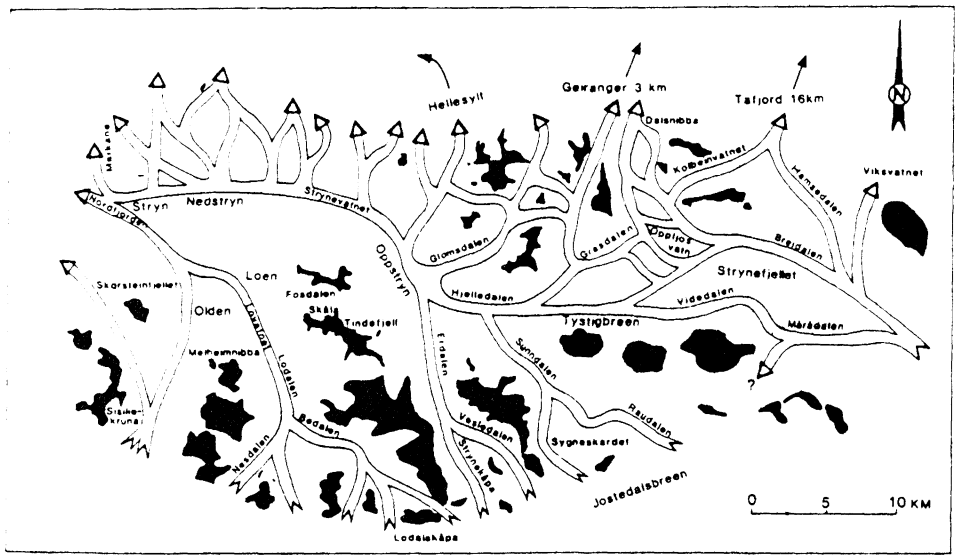


Fig. 14. Areas between Stryn and the northern Jostedal Plateau/ Strynefjellet covered by block fields (black areas), in some places with alpine morphology. The arrows indicate drainage routes for the inland ice during maximum glaciation. (Rye et al. 1987.)

The term block field is used for in situ angular blocks and stones formed through weathering of the local bedrock. The weathering limit is defined as the level between the lowest summits with block fields and the highest summits lacking block fields and having indisputable ice eroded forms. This limit has here a gradient of ≈ 7 m/km, and descends from about 1750 m a.s.l. at the northern Jostedal Plateau/Strynefjellet to approximately 1500 m a.s.l. in the mountain areas between inner Nordfjord and Sunnmøre (Fig. 15).

Rock surfaces between the Nor lateral moraines of Younger Dryas age (11 000-10 000 B.P.) and the weathering limit all show signs of glacial erosion, without any signs of block-field formation, despite the fact that they have been exposed to weathering from before 11 000 B.P. The weathering limit is up to 600 m higher than the Younger Dryas lateral moraines, indicating a more extensive glaciation. Rye et al. (1987) and Nesje et al. (1987) therefore concluded that the weathering boundary represents the upper level of the inland ice sheet during the last glacial maximum 18 000-20 000 B.P. It is possible, however, that the gently undulating mountain plateaus above the regional weathering boundary were covered by dynamically inactive local snow fields or minor ice caps too

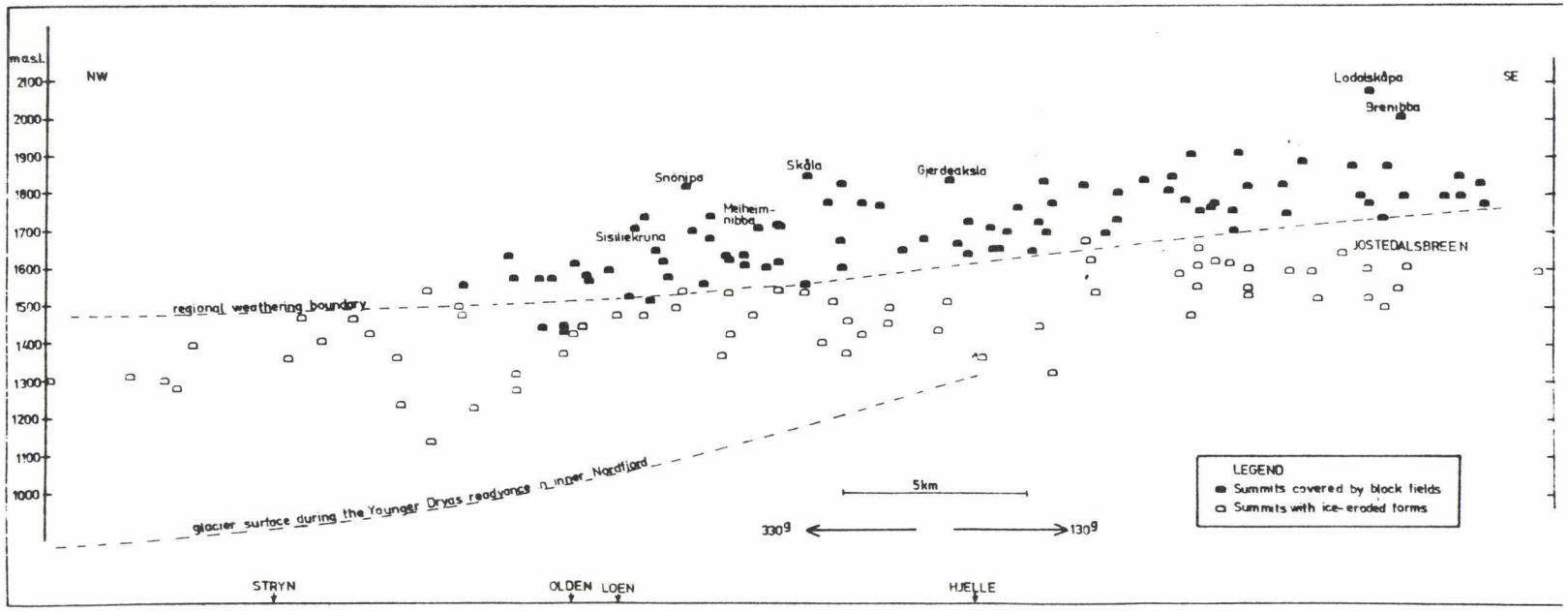


Fig. 15. Areas between Stryn and the northern Jostedal Plateau/ Strynefjellet with block fields, and summits with only glacially-sculptured surfaces plotted on a NW-SW profile, parallel to the main drainage direction for the inland ice during maximum glaciation. The elevation of the glacier surface in inner Nordfjord during the Younger Dryas is indicated.

thin to erode the block fields already formed.

A profile of the ice sheet during the last glacial maximum has been constructed (Fig. 16) by considering the weathering limit in inner Nordfjord, and the distribution of block fields, trimlines, erratics and glacial striae in the Møre area.

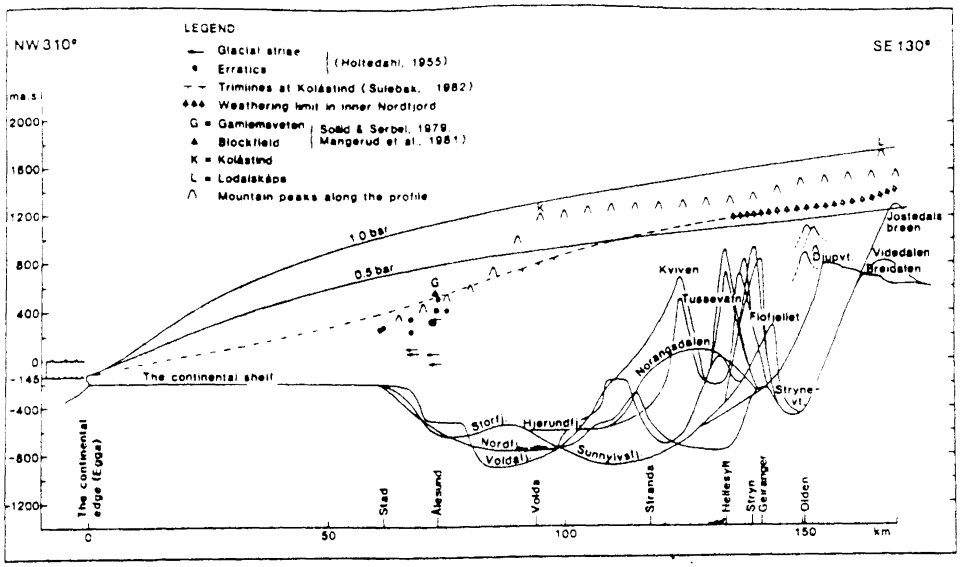


Fig. 16. Longitudinal profile from the continental edge NW of Møre to the northern Jostedal Plateau/Strynefjellet, showing the topography along the main drainage routes for the inland ice sheet. Theoretical glacier profiles for basal shear stresses of 1.0 and 0.5 bar are indicated (solid lines). The postulated glacier profile during Late Weichselian maximum is indicated on the basis of observed weathering limits in inner Nordfjord and observations from Møre (dashed line). The profile is corrected for a glacio-eustatic sea level 145 m below the present and a glacio-isostatic depression of c. 300 m in inner Nordfjord. (Nesje et al. 1987.)

Deglaciation

Fig. 17 shows reconstructed profiles of valley glaciers from the northern Jostedal Plateau/Strynefjellet during the Younger Dryas and Preboreal Chronozones.

The Nor moraines.

Cirque glaciers developed in the coastal area of Nordfjord beyond the terminus of the inland ice sheet (Larsen et al. 1984) during the Younger Dryas Chronozone. The most prominent center of local glaciation in Nordfjord was, however, the high-lying Devonian massif to the south of the fjord in the Gjegnalund/Ålfotbreen area (Fig. 13), where a large, continuous plateau glacier discharged

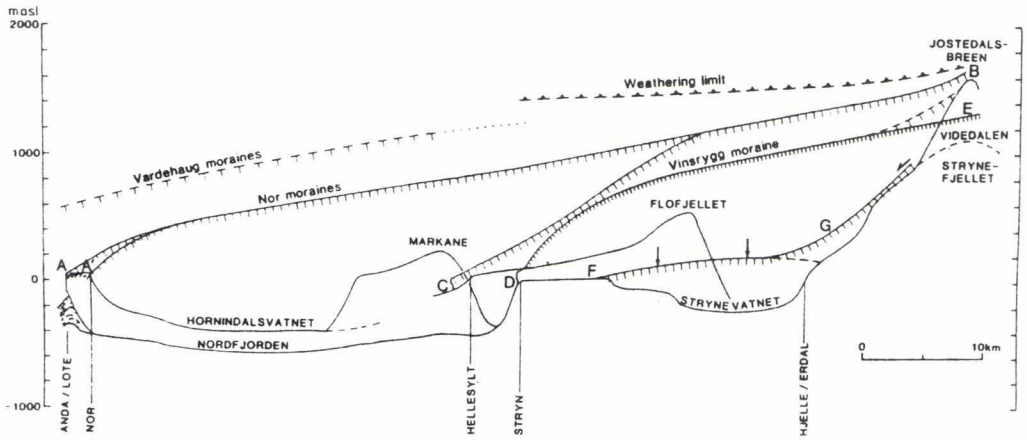


Fig. 17. Longitudinal profiles of valley glaciers from the northern Jostedal Plateau/Strynefjellet during the Younger Dryas and Preboreal Chronozones. The altitude of the weathering limit in inner Nordfjord and the glacier surface in the middle Nordfjord during formation of the Vardehaug moraine are also indicated: Profile A-B: Profile in Nordfjorden (the main fjord) from Anda/Lote to the Jostedal plateau during the Nor Stadial /Younger Dryas). Profile A'-B: Profile in the Hornindalsvatnet basin from Nor through the pass at Markane, west of Stryn, to the Jostedal Plateau during the Nor Stadial. Profile C-B: Profile during the Nor Stadial from an ice-front deposit in the inner part of Sunnylvsfjorden, through the pass at Flofjellet to the Jostedal Plateau. Profile D-E: Profile of the 'Stryn glacier' during the formation of the Vinsrygg moraines during the early Preborial Chronozone. (Rye et al. 1987.)

outlet glaciers into the surrounding valleys, some of them down to sea level. In addition, several cirque glaciers formed in the mountain areas of middle and inner Nordfjord above the outlet valley- and fjord glaciers descending from the Jostedal Plateau during that time (Rye et al. 1987, Fareth 1987). Their lateral moraines can be traced to the level of the Younger Dryas equilibrium line altitude (ELA) ≈ 450 m below the present-day level. Lateral moraines indicate that the glacier surface rose from an elevation of approximately 1100 m in the mountain areas at the valley mouths in inner Nordfjord to ≈ 1600 m a.s.l. at the northern Jostedal Plateau (Rye et al. 1987). This indicates a rather thin ice cover above the Jostedal Plateau. A branch of the "Stryn glacier" drained across the Flofjellet pass at 540 m a.s.l. to Hellesylt (Fig. 18).

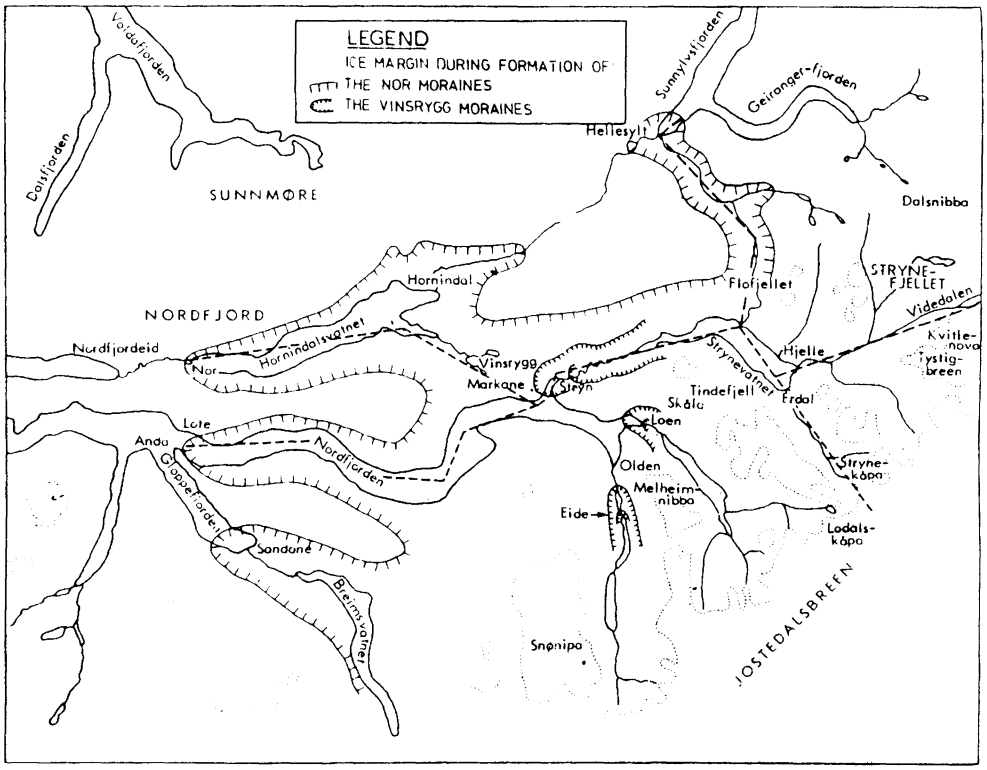


Fig. 18. Position of the ice margin in middle and inner Nordfjord during the formation of the Nor moraines. Positions of glacier termini in the inner Nordfjord during the formation of the Vinsrygg and Eide moraines are also indicated. (Rye et al. 1987.)

At Utvikfjellet (loc. 6), the Younger Dryas glacier in the main fjord (Nordfjord) received an additional supply of ice from the "Breim glacier". The surface of the latter lay some 200 m higher, causing the ice-flow to run northward in the pass area (pass elevation 640 m). The marginal belt commonly includes two-three, and in places four parallel ridges, which are generally continuous for long stretches.

The Vinsrygg- and Eide moraines.
The valley and fjord glaciers retreated rapidly from the position of the Nor moraines in middle Nordfjord to the valley mouths in inner Nordfjord some 40 km farther inland, caused by extensive calving of the fjord glaciers. Two ice-frontal deposits, only a few kilometers apart, are found in each of the valley mouths of the Stryn, Loen and Olden valleys, indicating minor readvances or at least halts in the general retreat of the glaciers (Fig. 18). The ice-marginal deposits are located where the fjord becomes shallower, commonly on rock thresholds and where the valleys are relatively narrow. The outer (distal) ice-frontal deposit in the main valleys are correlated morphostratigraphically with the Vinsrygg moraine, first described by Fareth (1970), with the type locality some kilometers west of the village of Stryn. The proximal deposits are correlated with the Eide moraine, the type locality being 3 km south of the village of Olden. Based on radiocarbon dates and marine levels, both the Vinsrygg and Eide moraines were deposited during the first half of the Preboreal Chronozone (10 000-9 500 B.P.).

The formation of the early Preboreal ice-frontal deposits may be explained in terms of glacial dynamic principles as a result of rapid calving of the fjord glaciers and unstable and dynamically active valley glaciers (Rye et al. 1987). During the succeeding stabilizing process, when the glaciers became grounded, the glaciers either readvanced or at least halted their general retreat, forming frontal deposits. Palynological data do <u>not</u> indicate a more severe climate during or soon before their formation.

A large boulder field occurs at the mouth of Myklebustdalen (loc. 7). It covers an area of about 1 km x 3 km. The field is interpreted to have slid down onto the surface of the valley glacier from the steep mountain walls bordering the southern part of Myklebustdalen and then carried by the glacier to the present position. It may thus be considered either a kind of ablation till, deposited in the marginal area of a slowly retreating glacier, or an end moraine.

The final deglaciation

After the deposition of the Vinsrygg and Eide moraines the sequence of deglaciation was characterized by downwasting glaciers in the inner areas of the Stryn and Loen valleys, and a retreating valley glacier in the Olden valley.

At Hjelle is a large ice marginal deposit, Hjellevonga (loc. 4), formed of a downwasting valley glacier. It has a weakly undulating surface about 75 m a.s.l., indicating that it was built up to sea level at that time. The steep western slope toward Strynevatnet may be primarily ice-contact formed at the margin of a stagnated glacier in the Strynevatnet basin.

In this period the glacier and meltwater drainage in inner Nordfjord was controlled by mountain passes at Strynefjellet (Fig. 13), which became ice-free in middle Preboreal.

The valleys surrounding Jostedalsbreen were deglaciated during the later half of the Preboreal Chronozone (9 500-9 000 B.P.). However, a glacier readvance followed at the end of the Preboreal, with the belonging terminal moraines located ≈ 1 km beyond the Little Ice Age moraines. These are found in several valleys around Jostedalsbreen. At Vetledalssetre in Erdalen (loc. 3) is a 500 m long, blocky end moraine, which reflects a climatically-controlled readvance of glaciers on the Jostedal Plateau during the latest part of the Preboreal Chronozone. A minimum age of 8 810±130

radiocarbon yrs B.P. (Nesje 1984), was obtained from a 4 m deep peat bog 300 m proximal to the moraine.

Time-distance diagram

The most recent time-distance diagram for the deglaciation of Nordfjord is that proposed by Rye et al. (1987) (Fig. 19). The geographical and in part chronologial extent of the local glaciation is schematic. There may be large local differences in deglaciation times along the length of the fjord, as illustrated by the difference in time of deglaciation in the adjacent Olden and Stryn valleys.

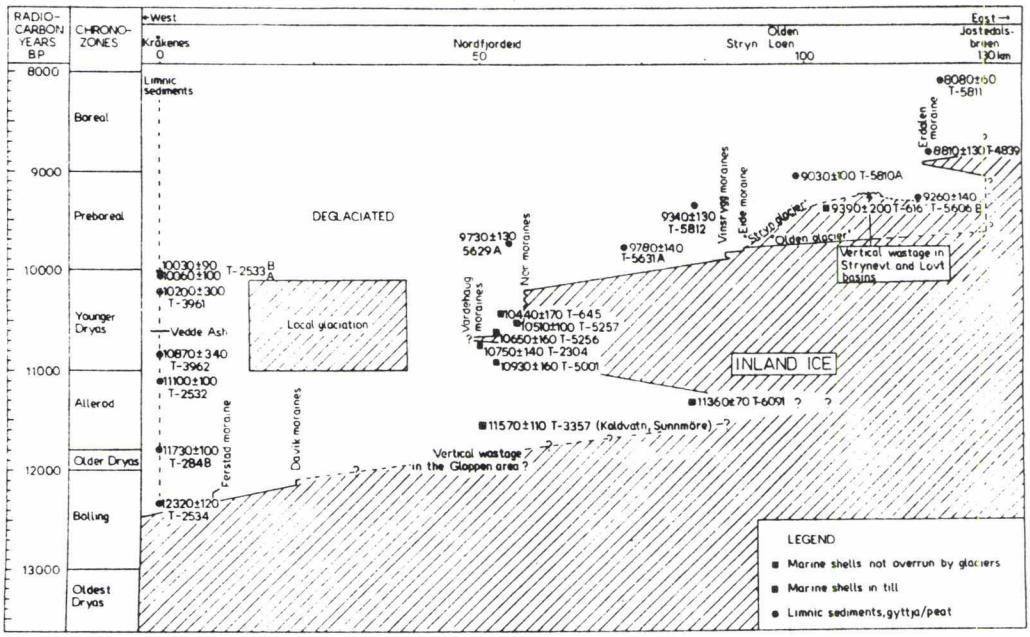


Fig. 19. Time-distance diagram for the deglaciation of Nordfjord. (Rye et al. 1987.)

The Holocene glacial and climatic development in the Jostedalsbre region

The equlibrium line altitude (ELA) was ≈ 350-400 m below the present during the late Preboreal glacier advance. Assuming similar precipitation to the present, this suggest a mean temperature

decline of ≈ 2.0-3.0 ^{0}C. Palynological investigations from Sygneskardet, Sunndalen indicate that climate like the present occurred just after 9 000 B.P. The Holocene climatic optimum in the Jostedalsbreen region occurred between 6 500-5 000 B.P. with elm (Ulmus) growing at the present birch tree limit about 700 m a.s.l. in Sunndalen and Pine (Pinus) growing at Styggevatnet to an altitude of at least 1160 m a.s.l. The mean summer temperature is estimated to have been 2.5 ^{0}C warmer than at present during the mid-Holocene climatic optimum. An inferred rise of the ELA of ≈ 400 m from the present altitude (1500-1550 m a.s.l.) suggests that only a few, or possibly no glaciers existed then at the Jostedal Plateau.

Lowered tree limit and increased resedimentation into peat bogs indicate climatic deterioration from 5 000 B.P. The first Neoglacial phase in the Jostedalsbre region was bracketed between 3 700-3 100 radiocarbon yrs B.P. Thereafter the glaciers retreated, and the present glaciers were formed subsequent to 2 000-1 500 B.P., or possibly during the initiation of the Little Ice Age (≈ AD 1 350).

We have dated gelifluction at the altitude of 1000 m a.s.l. to have initiated close to 3 000-2 800 B.P. The glaciers were smaller than at present during the Medieval period.

A severe climatic deterioration took place from the 14th, and especially from the mid-17th century. This culminated in the mid-18th century during the peak of the Little Ice Age, and is historically documented. Documents indicate that the western outlet glaciers from Jostedalsbreen reached their maximum Little Ice Age position some years before the longer eastern outlet valley glaciers. A 150 m lowering of the ELA during the Little Ice Age, inferred from the upper level of lateral moraines, suggests a mean annual temperature ≈ 1 ^{0}C lower than at present. Fig. 20 indicates the fluctuations of the ELA and the inferred mean paleotemperature variations throughout the Holocene.

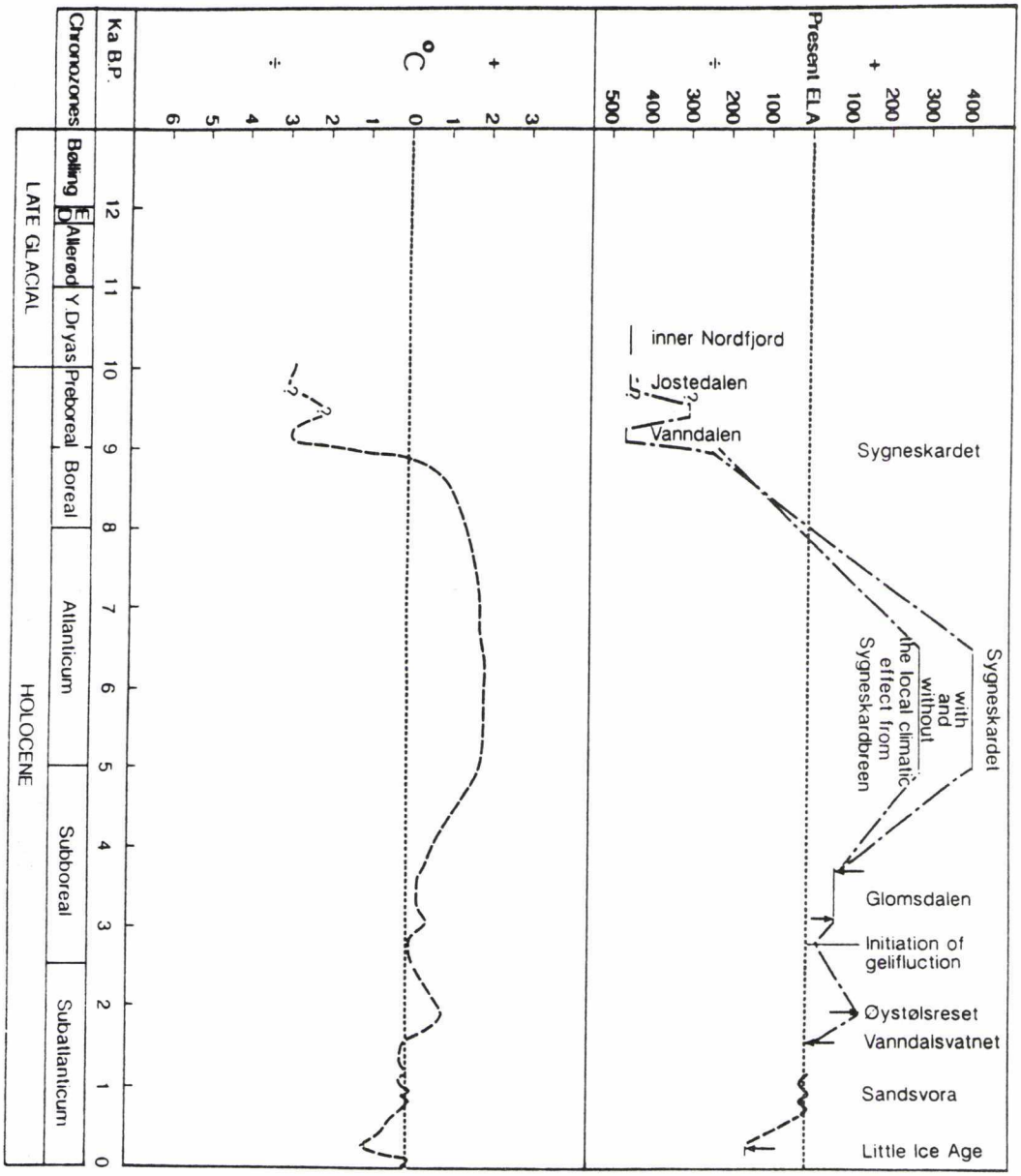


Fig. 20. Fluctuations of the ELA and the inferred mean palaeotemperature variations throughout the Holocene in the Jostedalsbre region. (Nesje et al. in prep.)

Bødalen (loc. 2) contains several little Ice Age moraines deposited up to 1.6 km in front of the present position of Bødalsbreen (Fig. 21). The outermost moraine has the "normal" curved form of terminal moraines, but inside this are unusual saw-tooth-shaped moraines. Lien and Rye (1988) explain these as being push moraines. The shapes are caused by valley widening at this locality, leading to lateral extension and development of wedge-shaped crevasses which gave the ice front a saw-tooth appearance.

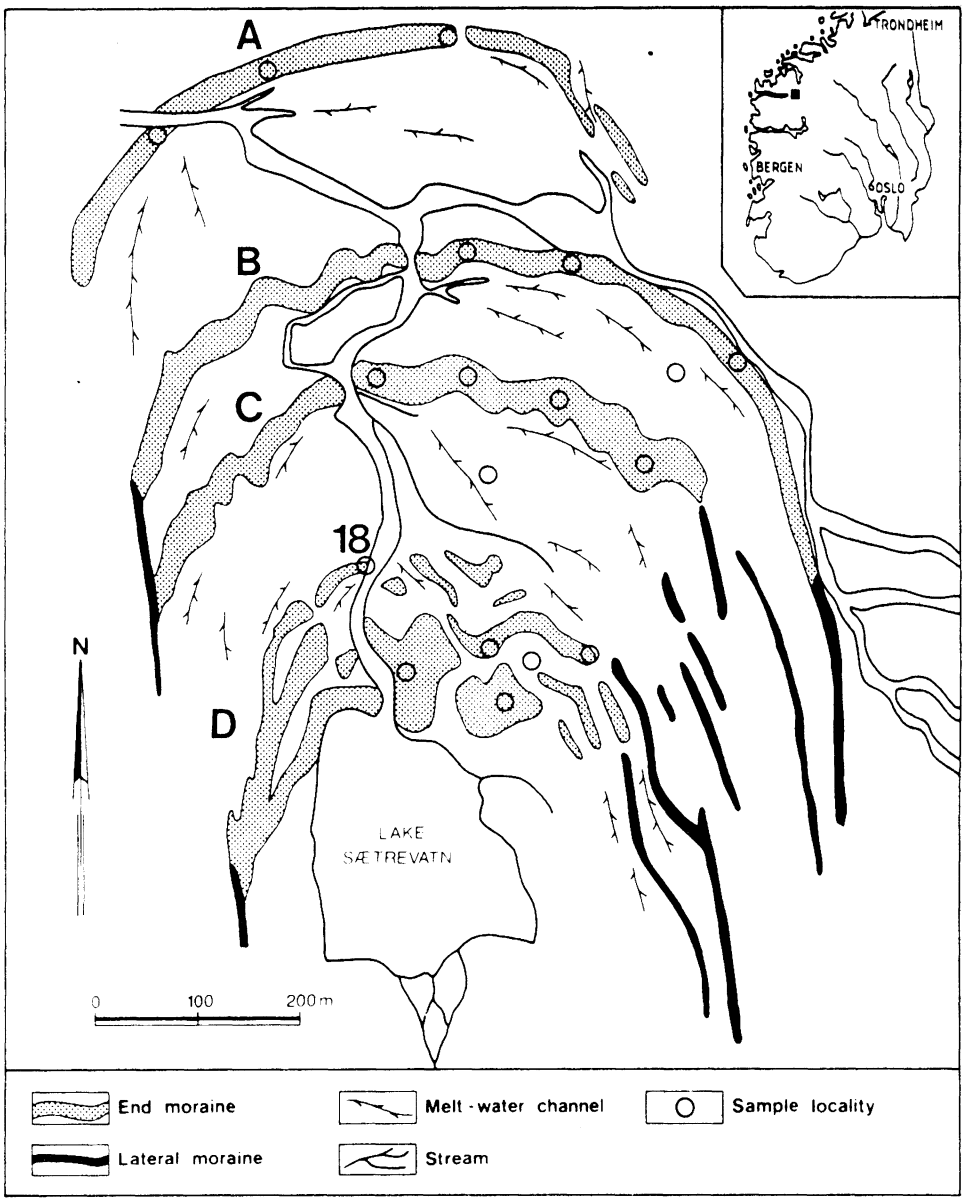


Fig. 21. Little Ice Age moraines in front of the Bødalsbreen glacier, western Norway. (Lien & Rye, 1988.)

HOLOCENE CHANGES IN JØLSTRAVATNET

The lake Jølstravatnet (loc. 8), situated at the watershed between Nordfjord and Førdefjorden (Fig. 22) demonstrate the effect of glacio-isostatic rebound on lake development, because of the small difference in elevation of the thresholds at both ends and because the isobases are crossing the lake at an oblique angle.

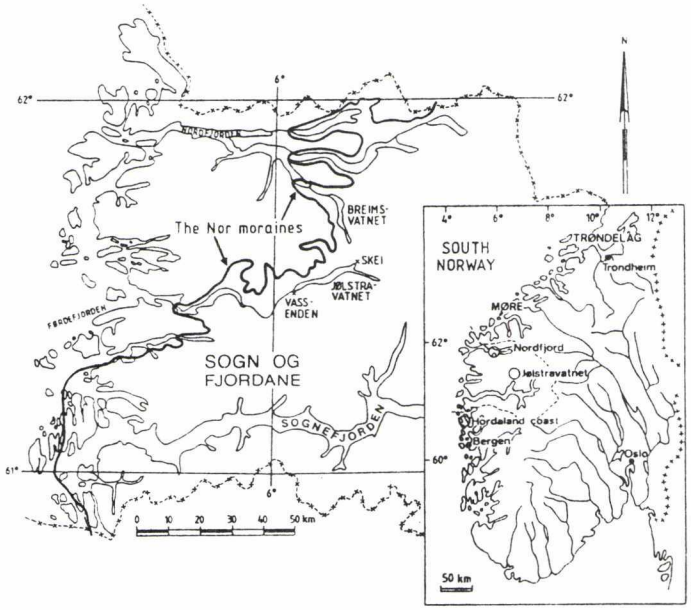


Fig. 22. A map of the area showing Jølstravatnet situated just inside the Nor moraines. Inserted is a key map of southern Norway.

Jølstravatnet is 29 km long. Its present outlet lies at Vassenden (Fig. 23). The 80 m broad outlet river (Jølstra) passes across a bedrock threshold which has a veneer of boulders. The level of the threshold is estimated to 205.9 m a.s.l. from the lowest recorded water level in the lake. A threshold at the northeastern end of the lake (at Skei) is only ≈ 3 m higher than the present outlet threshold.

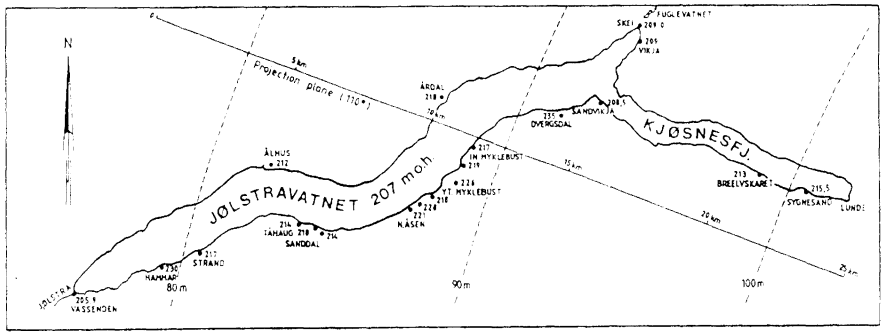


Fig. 23. Terrace and threshold levels at Jølstravatnet. Isobases for the Nor level (Younger Dryas marine shore level) are from Fareth (1987).

The lake area was deglaciated during the Preboreal. After the formation of the Nor moraines in Younger Dryas both the lobe in Førdefjord and in Nordfjord (Fig. 22) retreated to this area. The divide between the two lobes was situated at the watershed at Skei.

The lake history can be subdivided into five phases (Fig. 24). Approximate 14C ages for the phases are obtained by the construction of a time-gradient curve for the area. During phase I (-9 500 B.P.), Jølstravatnet was an ice-dammed lake. Phase II (9 500-9 200 B.P.) started with the burst of the ice-dammed lake and a 12 m lowering of the lake level. An eastern outlet of this new glacial lake was established. During phase III (9 200-7 500 B.P.) glacio-isostatic rebound caused a fresh water transgression in the western part of the lake. Phase IV (7 500-6 000 B.P.) is characterized by outlets at both ends of the lake. The isolation of the eastern outlet (≈ 6 000 B.P.), marks the transition to phase V, i.e. the present situation with outlet over the western threshold.

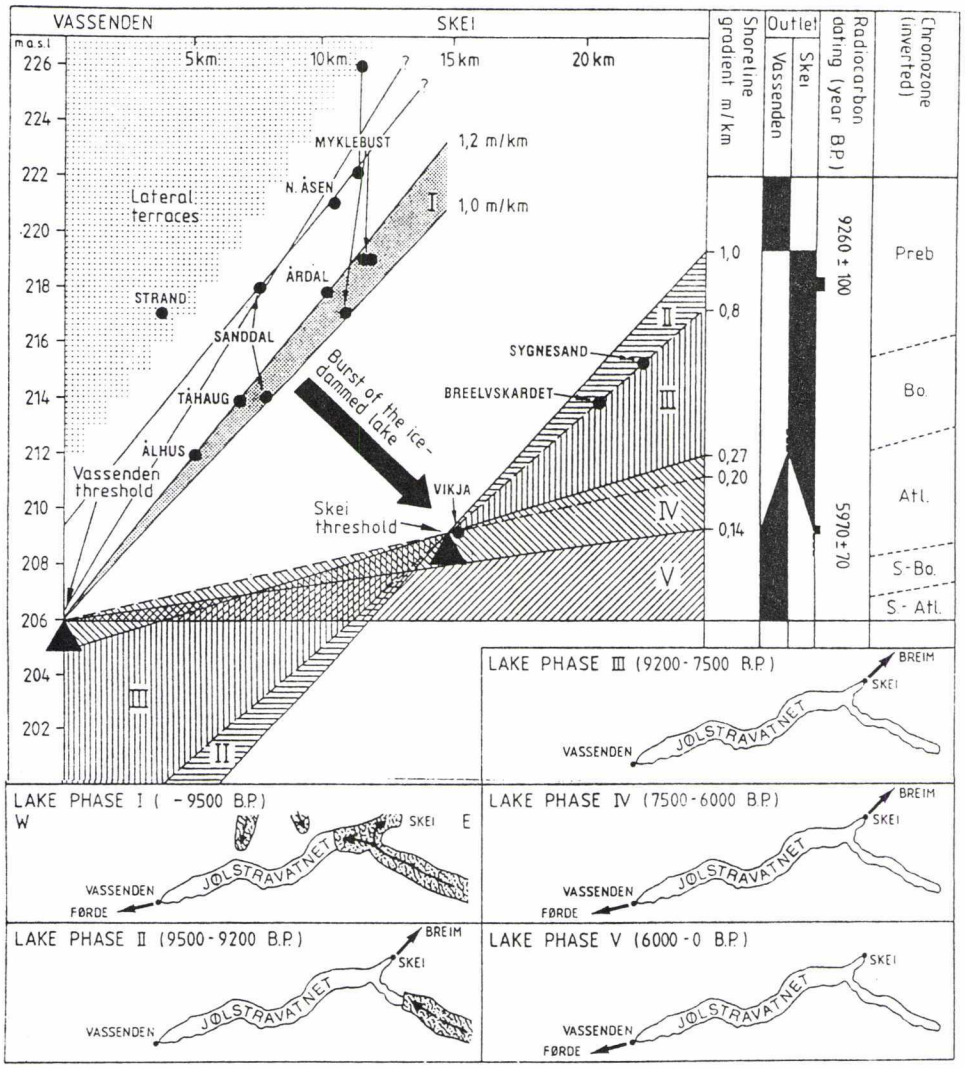


Fig. 24. The equidistant shoreline diagram for Jølstravatnet and phases in the Holocene changes of the lake. (Note that the stratigraphic columns are turned upside-down in order to match the diagram.)

GLACIOLOGY AND GLACIAL GEOLOGY OF FJÆRLAND

Introduction

More than 1% of the area of mainland Norway is covered by glaciers. Jostedalsbreen has an area of 486 km^2 and is the largest glacier on the continent of Europe. Adjoining glaciers cover an additional 330 km^2. More than 30 named outlet glaciers of Jostedalsbreen spill down into the low-lying valleys. The higher glaciers were believed to cover fairly level bedrock at high elevations. However, drillings and radioecho soundings have now shown that the subglacial topography is not a level plateau and that Jostedalsbreen is up to 600 m thick.

Glaciers cover many of the mountains surrounding Fjærland. Two of the branches from Jostedalsbreen, Bøyabreen and Supphellebreen, are the most prominent. Both these glaciers reached their maximum post-glacial extent around 1750 A.D. The advance of Supphellebreen destroyed good farmland, but there are no records in Fjærland of measurements of glacier position from that time. Historic records elsewhere around Jostedalsbreen show that a farm was destroyed by Nigardsbreen in 1742, and another by Åbrekkebreen in 1743. Measurements were thereafter done at Nigardsbreen, because the King wanted recorded evidence in connection with petitions from the farmers requesting reduced taxes because of land destruction. This glacier reached its maximum position in 1748. The first studies of Bøyabreen and Supphellebreen were done in the late 1860's by Sexe and de Seue.

Measurements of front positions of outlet glaciers of Jostedalsbreen were initiated late last century by Rekstad. They were mostly organized by Bergen Museum in the first half of this century, and thereafter by Norsk Polarinstitutt. Position has been measured of 14 of the outlet glaciers from Jostedalsbreen, mostly by persons living near the glacier. Today measurements continue only at six glaciers. This reduction is partly because Norwegian studies on glaciology/climate have increasingly concentrated on mass balance measurements, and partly because the distance from the

original markers to the present glacier front have become very large, and new markers have not been established. Fig. 25 and Fig. 10 (in Bogen, this volume) show length variations of various outlet glaciers discussed by Liestøl (1967) and Bogen, Wold and Østrem (1988). The former computed the mass balance variations using data on winter precipitation and summer temperature recorded in Bergen to represent accumulation and ablation, and Liestøl showed that the glacier fluctuations are very well reproduced with a time lag of four years.

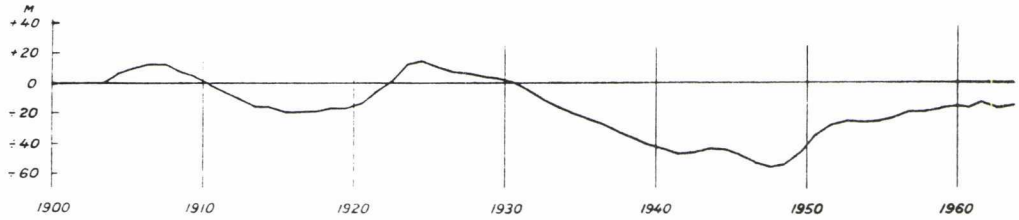


Fig. 25. Annual front variations of the outlet glaciers from Jostedalsbreen, smoothed to a four-year mean (from Liestøl 1967, Fig. 24).

Pedersen (1976) combined discharge measurements for Lovatnet and Oldevatnet with glacier front measurements to compute annual variations in glacier mass. She showed that the glacier area of this part of Jostedalsbreen decreased from 216 km^2 in 1903 to 154 km^2 in 1974.

The excursion will pass through the moraine system of Bøyabreen (loc. 9, below), and it is planned to visit the upper part of Supphellebreen on the following day. The lower parts of the glaciers may be visited by those not taking part in the high excursion, and are described below. Additional low-lying localities related to glacier erosion and deposition may be visited if conditions hinder the high excursion. Some of these localities are discussed in Mundal (1953).

Bøyabreen

The road ends at the 1930 moraine at an elevation of 150 m. One should not walk the <1 km distance to the glacier because of rock avalanche danger. 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 a probably ≈ 100 m thick.

Bøyabreen was one continuous glacier when the moraine was deposited in 1930 and one could easily walk onto the glacier tongue. The terminal moraine from ≈ 1750 is located by the sæter huts, about 2 km further down valley. The whole moraine system (loc. 9) is a typical example of the systems formed during the past 240 years in front of the outlet glaciers from Jostedalsbreen. It was studied by Fægri (1933), who used the plant succession to derive relative ages of the moraines (Fig. 26), and compared these with the moraine systems in front of Nigardsbreen, Bersetbreen and Åbrekkebreen. Front positions of Bøyabreen were measured from 1903 to 1952.

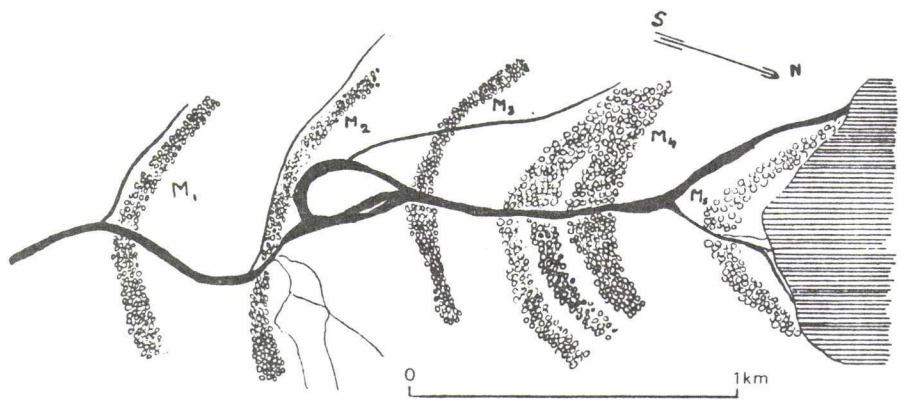


Fig. 26. The moraine system in front of Bøyabreen in 1931, from Fægri (1933, Fig. 44). He gave the moraines M_1 to M_4 estimated dates of 1750, 1825, 1840, and 1875 respectively, and M_5 was observed to 1909.

Lower Supphellebreen

The road leading up to the lower, regenerated, glacier ends also here at the terminal moraine from 1930. The altitude is only 55 m, making this the lowest glacier in the hemisphere south of the Arctic circle. The glacier is only about 600 m long, but reaches to 300 m elevation. The front of the upper glacier is at 700 m elevation. Here it has a flow rate ≈ 2 m/d and is ≈ 50 m thick. Blocks of ice break off where the slope becomes too steep, and they cascade over the cliff and are broken into small fragments before settling on the lower glacier.

Orheim (1970) mapped the glacier in spring and autumn from 1964 to 1967 and showed that about 2 million tons of ice fell every year from the upper glacier to nourish the regenerated glacier. This was

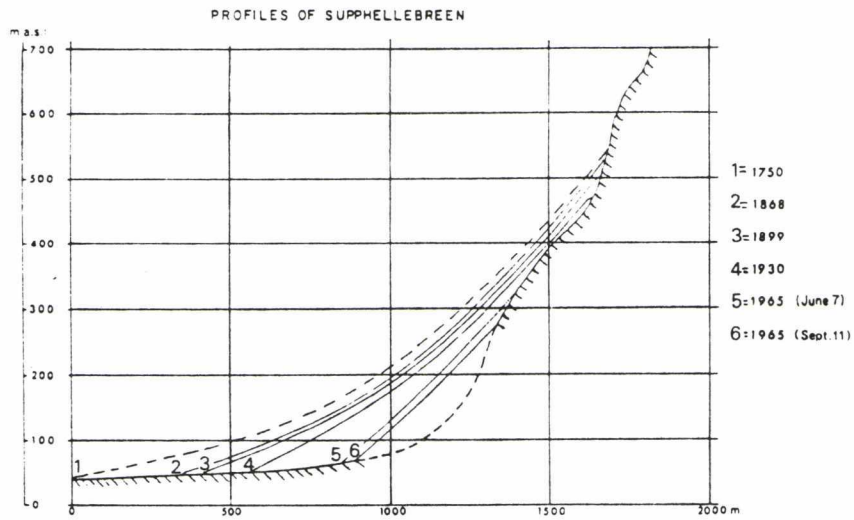


Fig. 27. Profiles of lower Supphellebreen from ≈1750 to 1965. (From Orheim 1970.)

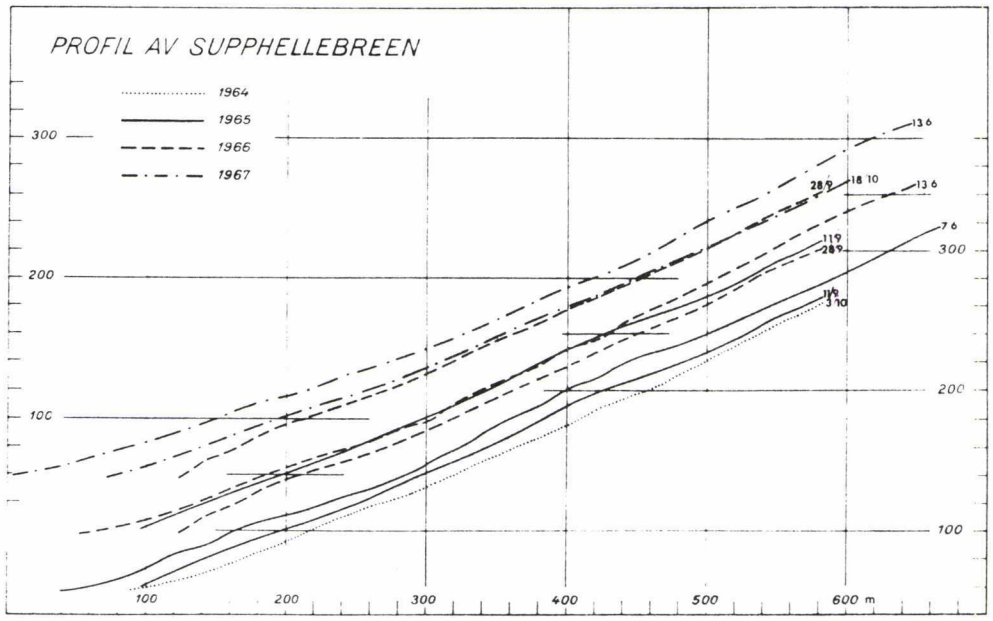


Fig. 28. Seasonal variation in thickness of lower Supphellebreen from 1963 to 1967. Note that the individual years are transposed 40 m along the vertical, and that profiles for September are repeated.

equal to adding a 15 m thick ice layer to the lower glacier (Figs. 27 and 28), 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!

Most of the ice falls in the winter, when there can be several ice avalanches every hour. Calvings are frequent also in May and June, and are occasionally observed later in the summer. An ice fall will be heard as a loud, thundery noise, and comes as a stream of broken ice, which from below looks like a waterfall. An avalanche may last up to 1 minute. The lower Supphelle glacier will look very dirty in September, because rock material that falls from above accumulates on the surface, and older rock fragments may also become exposed.

Front positions of Supphellebreen were measured from 1901 to 1958, and from 1964 to the present. Position of boulders marked in 1964 and 1976 show that the glacier dimensions have changed little in recent decades. Fig. 29 shows position of the glacier from 1750 to present, from Orheim (1970).

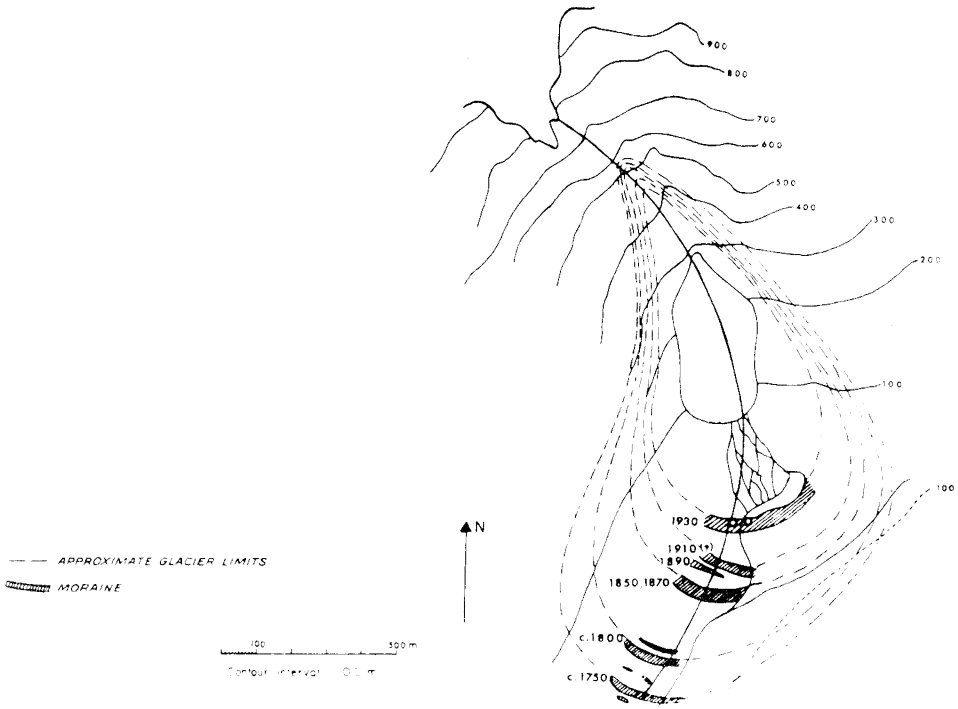


Fig. 29. Extent of lower Supphellebreen from ≈1750 to 1930. (From Orheim 1970.)

Upper Supphellebreen

The mass balance of the complete Supphellebre glacier system was studied 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, by 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. 30). Data are tentative for five years, and

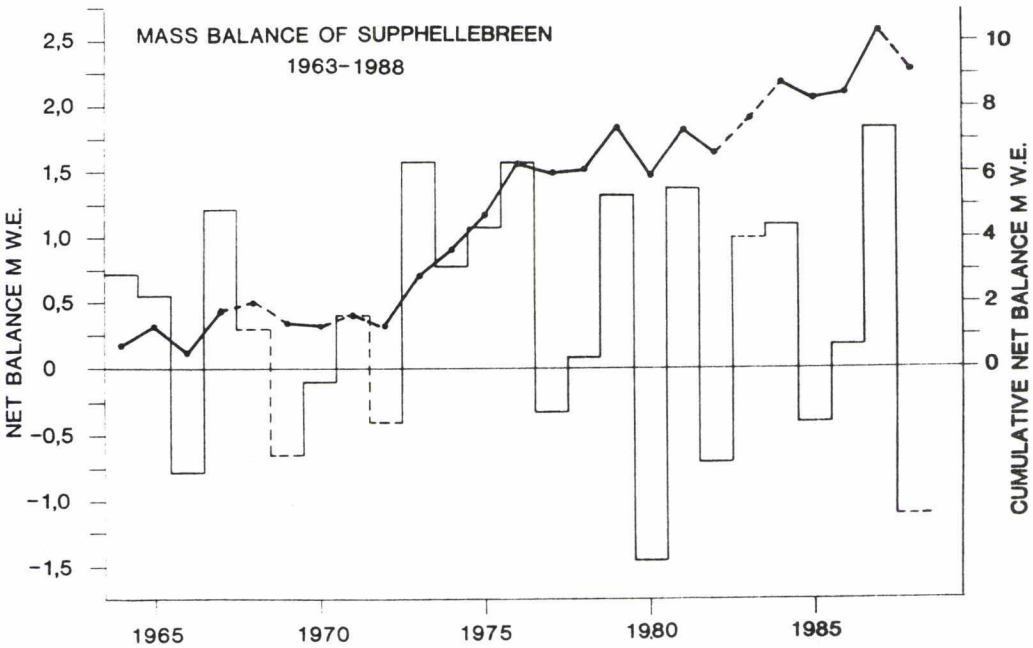


Fig. 30. Annual and cumulative net mass balance of Supphellebreen from 1963 to 1988. The measurements include two-year-means for 1968 and 1969, and 1971 and 1972, and the annual balance for these four years were estimated by comparing with Nigardsbreen. The value for 1983 is also estimated from Nigardsbreen data. The estimated 1988 value is based on measurements up to August 1988.

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 that the front of Supphellebreen has only advanced a few m between 1968 and 1988.

A lake is dammed between the complex terminal moraine of Flatbreen (Supphellebreen) and the glacier. The lake is filled each spring, and it drains under Supphellebreen during the summer. The time of drainage has advanced from August/September in the 1960's to July this year probably because the glacier has thinned. The drainage occurs over a short period of time (hours), in a manner described for similar lakes by Liestøl (1956).

SOGNEFJORD

The Sognefjord is the largest fjord system in Norway, penetrating about 170 km into the heart of the country. It is also the deepest, having a maximum depth of 1308 m. The outer part of the fjord has a few tributary fjords, all of which are short and bluntheaded. The inner part, however, has a number of extensive branches, such as Nærøfjord, Aurlandsfjord, and Fjærlandsfjord. The height of the surrounding mountains increase from west to east, being about 500 m in the outer area near the mouth of the fjord, and 1100-1200 m in the area of Aurlandsfjord. Near the coast the fjord is bounded by low islands and skerries which are part of the Norwegian strandflat. The low ledges along the shores of the fjord, best developed in the outer fjord area, are by Ahlmann (1919) and Nansen (1921) thought to represent the remains of a continuation of the strandflat. They are assumed by the former to represent the remains of an old pre-glacial valley generation, partly worn away by the erosional effect of the inland ice, and by the latter to have been formed by shore erosion and frost action during the Pleistocene.

The Sognefjord cuts through a part of the north-western gneiss area with east-west, northeast-southwest structures of Precambrian origin, and then penetrates the Caledonian fold-trough in the east. There is no obvious relationship between the trend of the fjord and the fold pattern, with exception of some of the branches at the inner part of the main fjord. There is, however, a relationship between fracture systems and the trend of the main fjord and its branches, and these fracture systems have played an important part in guiding erosion.

A very marked north-south fracture pattern is present at the mouth of the fjord, especially well developed in the Devonian rock complex.

The Sognefjord as a glacially overdeepened rock basin

A longitudinal section of the Sognefjord (Fig. 31), as well as transverse sections, show features regarded as typical of a fjord: a basin-like form, with generally steep sides, and a partly flat bottom. The main fjord, which may be regarded as beginning with Årdalsfjord or Lærdalsfjord, increases evenly in depth westwards, so that it gradually passes into the extensive central basin with depths of 1000-1300 m between Leikanger and Brekke. From Brekke the floor rises rapidly to the area south of the Losna island, where a threshold extends westwards with depths varying between 100 and 550 m.

The great threshold west of Brekke is no doubt a rock threshold, and consequently the Sognefjord must be regarded as a rock basin eroded and overdeepened by glacial erosion.

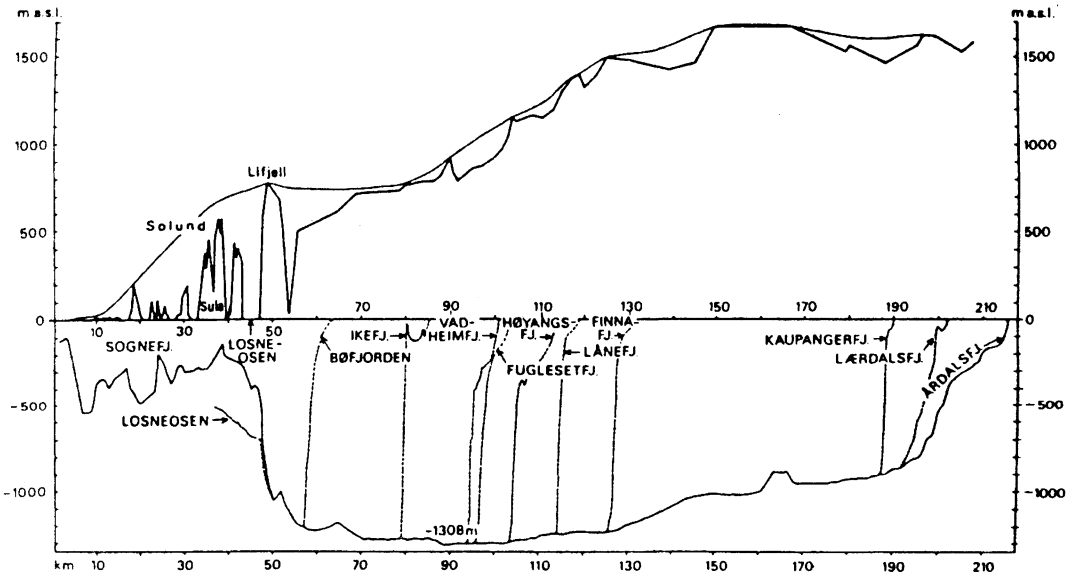


Fig. 31. Longitudinal profile of Sognefjord and general level of the area on the northern side of the fjord. Dotted lines: tributary fjords.

The tributary fjords of the Sognefjord are all hanging in relation to the main fjord, frequently with a basin-like form and threshold at their outer end.

It is generally thought that the major outlines of the Sognefjord system to some extent reflects a preglacial valley system (Holtedahl 1975). A reconstruction of these palaeic valleys has been attempted by several authors, but must be regarded as highly hypothetical, as the effects of glacial erosion during the entire Quaternary period must have been substantial. Remnants of a paleic valley system in the inner Sognefjord area are thought to be present in the Nærøfjord - Nærøvalley ledges about 500 m a.s.l., increasing in height towards the northeast. The valley and fjord below this level are no doubt mainly formed by glacial erosion. The back-cutting effect of valley glaciers is shown up clearly in the steep valley head at Stalheim.

SEDIMENTS AND SEDIMENTARY PROCESSES IN THE FJÆRLANDSFJORD AND PARTS OF THE SOGNEFJORD

The following is based on unpublised theses in marine geology, unpublished seismic profiles from the Sognefjord and various published data on the Quaternary geology of the area.

From the general geomorphology of the fjords (see Holtedahl this volume) we realize that they comprise a series of well developed sedimentary basins. The 1300 m deep basin in the Sognefjord, as well as many of the basins in the tributary fjords, have sediment thicknesses of 200-300 m. The major part of the sediment input to the fjord basins took place during the deglaciation of the Late Weichselian ice sheet.

The Sognefjord

In contrast to the Nordfjord area (Rye and Nesje this volume) the entire Sognefjord was filled with ice until the later part of the Younger Dryas Chronozone. Terminal moraines were formed at the entrance of the fjord (Aarseth and Mangerud 1974), where the water depth is in the order of 200-400 m (see profile in Holtedahl this volume). As the ice front retreated beyond the sill area, the fjord glacier calved back rapidly to the next standstills at the heads of the tributary fjords where glaciofluvial ice front deltas were formed some 500 years later (Vorren 1973; Bergstrøm 1975). During this short time span, the glacier front in the fjord most probably calved back at an average rate of some 400 m/year.

Each part of the Sognefjord basin was in a proximate position to the front of the main fjord glacier for a relatively short period, and the bulk of the suspended matter was deposited close to the point where the fresh meltwater met the saline fjord water. After the deglaciation of the main fjord, the glaciers in the tributary fjords still fed the main fjord with meltwater and sediments through most of the Preboreal Chronozone (10 000-9 000 B.P., Aarseth 1980). The sediment supply from these lateral sediment sources increased when the tributary fjord basins were filled to a

level higher than the rock thresholds. Instability then caused slumping of great quantities of sediments down to the main fjord basin. One example of this is where the Fjærlandsfjord enters the Sognefjord (see below).

The following localities are described along the excursion route Fjærland - Flåm.

Inner Fjærlandsfjord

Due to a very rapid glacial retreat in the inner part of the Fjærlandsfjord, the sediment thickness is only 40 - 80 m, (Fig. 32, Vangsnes 1981). On the delta and in the talus area at Berrføtlene it may exceed 100 m. The longitudinal profile of the fjord comprises two basins and two thresholds in addition to the delta and prodelta areas. The main sources of sedimentation are at present glacial flour from some of the outlet glaciers of the Jostedal ice cap brought to the Bøya delta, and various lateral sources along the entire fjord, such as tributary rivers and brooks, snow avalanches and rock falls. On the basis of this, three

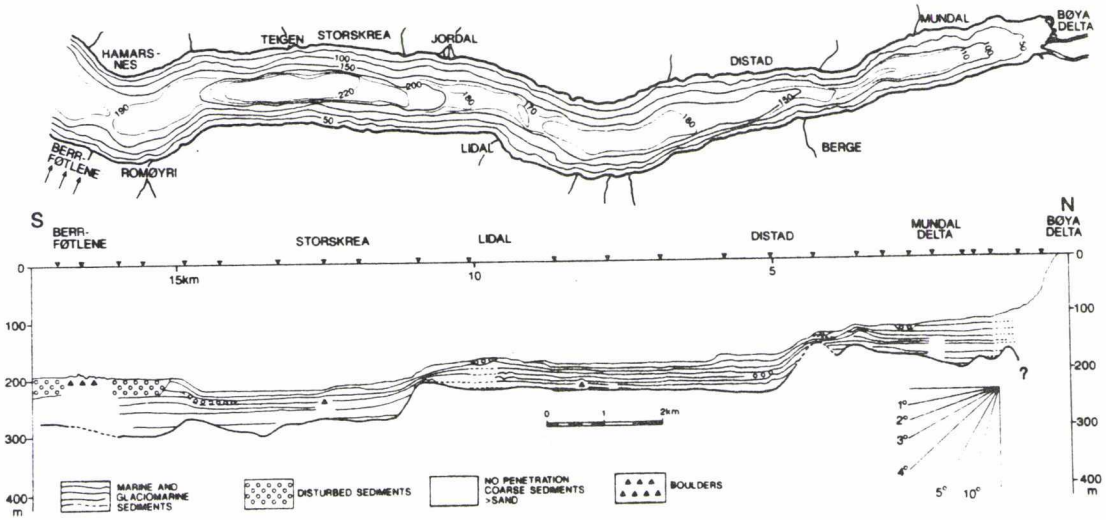


Fig. 32. Contour map and longitudinal profile of the sediment distribution of Inner Fjærlandsfjord, (modified from Vangsnes 1981).

different sedimentary facies can be defined along the fjord: delta-, basin- and threshold-facies, with various input from the two main sediment sources. This is well demonstrated by the different grain size of the surface sediments along the fjord. The silt ratio is rather uniform (approximately 40%) except in the prodelta area where it may exceed 60%. The sand/clay ratio varies with approximately 50% clay and <10% sand in the basins, and <30% clay and approximately 25% sand on the thresholds, (Fig. 33, Vangsnes 1981).

The recent sedimentation rates in the inner part of the fjord was measured by the ^{210}Pb-method in two sediment cores. Outside the Mundal delta, 2.2 km from the head of the fjord this method gave a mean value of 4.0 mm year^{-1} over the last 90 years, while the values 4.4 km further out in the fjord gave only 1.2 mm year^{-1} for the last 45-50 years (Vangsnes 1981). According to Mundal (1953) the delta front prograded with an average rate of 2.5 m year^{-1} in the period 1830-1950.

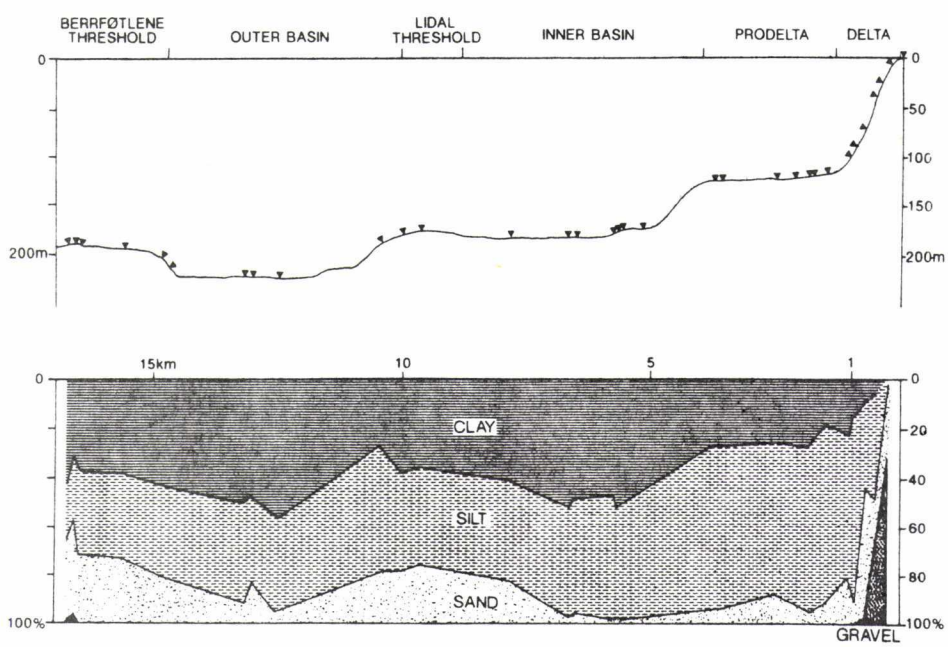


Fig. 33. Variation of the grain size in the surface sediments of the Fjærlandsfjord. Dots along the profile mark sample locations, (Vangsnes 1981).

The terrestrial slope processes are obvious along the fjord sides of the Fjærlandsfjord. Old rock avalanche deposits are recognized in seismic records of the fjord sediments (Fig. 34). The rock material have moved almost right across the horizontal sediment bottom (Vangsnes 1981). The largest talus area along the fjord is located at Berrføtlene (loc. 11), just outside Romøyri, (Fig. 32). Here, a 800 m high and 1500 m broad talus cone continues as a submarine accumulation on the 200 m deep fjord bottom. Side-scan sonar records together with acoustic reflection profiles give good picture of this accumulation (Fig. 35). The source for this talus is a heavily fractured bedrock zone approximately 1000 m a.s.l. clearly visible on the southwestern mountain side. The mass movements are still very active, with wet snow avalanches being the main transporting agent in addition to frequent rockfalls. A plunge pool as described by Liestøl (1972) is found where the path of the most active snow avalanches hits the fjord. The form of the submarine accumulation, however, suggests that larger rock slides must have occurred sometimes in the past.

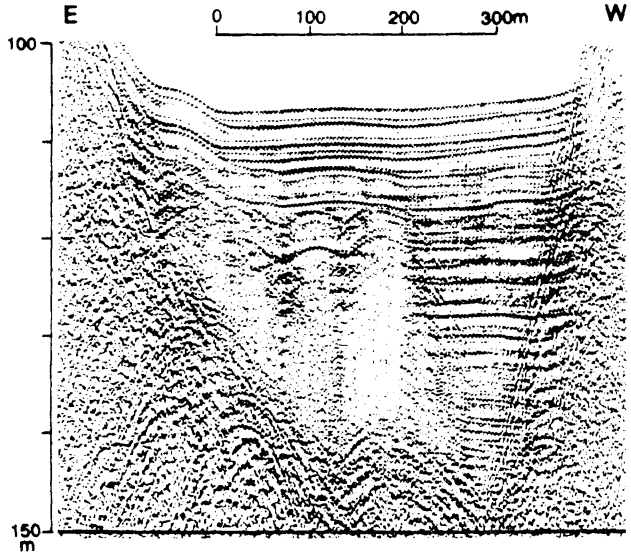


Fig. 34. Seismic profile across the fjord at Storskrea, Fjærlandsfjord, (Vangsnes 1981). For location see Fig. 30.

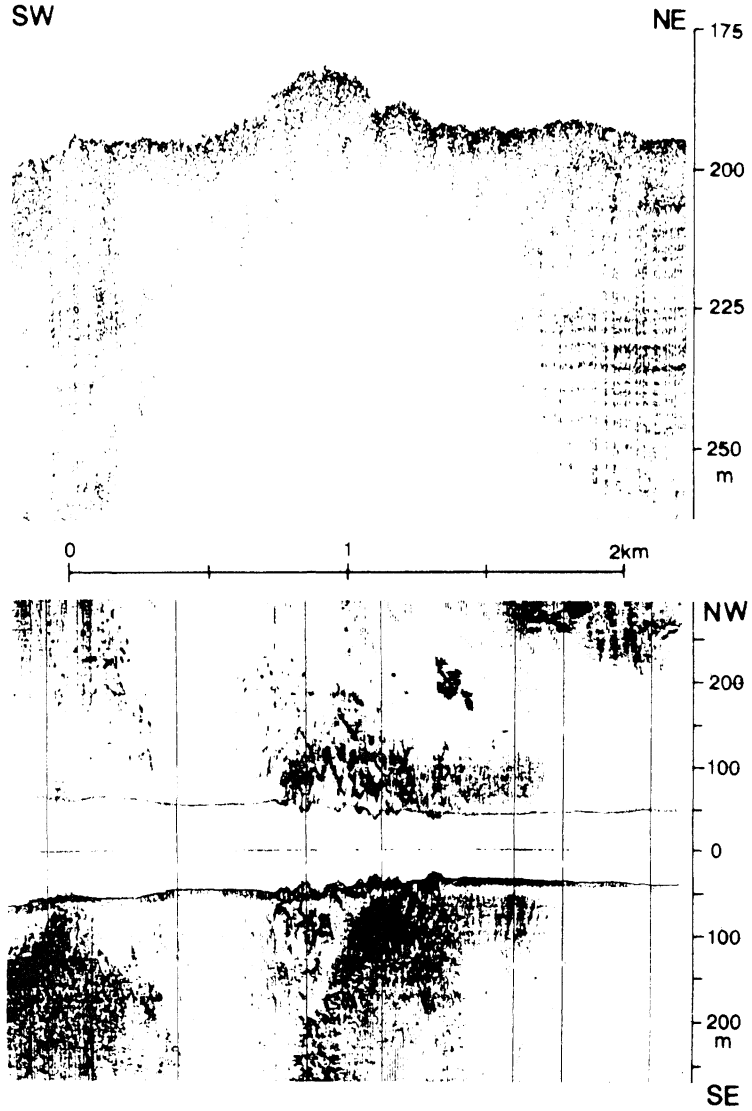


Fig. 35. 3.7 kHz bottom profile (upper) and side scan sonar record (lower) of the fjord bottom at Berrføtlene (Fig. 30). The side scan record has the ship track in the middle and "sees" to both sides. The rock avalnches from Berrføtlene on the southeast side has passed the middle of the fjord and can be traced as boulders on the northwest side of the fjord bottom.

Outer Fjærlandsfjord - Vetlefjord

The Fjærlandsfjord is a typical "hanging fjord" where it merges with the Sognefjord at Hella (Fig. 36, Lønne 1981). The calving activity decreased during the deglaciation, and more sediments were deposited on the fjord bottom. Fig. 37 indicates two areas of glaciofluvial foreset beds and up to 200 m of glaciomarine

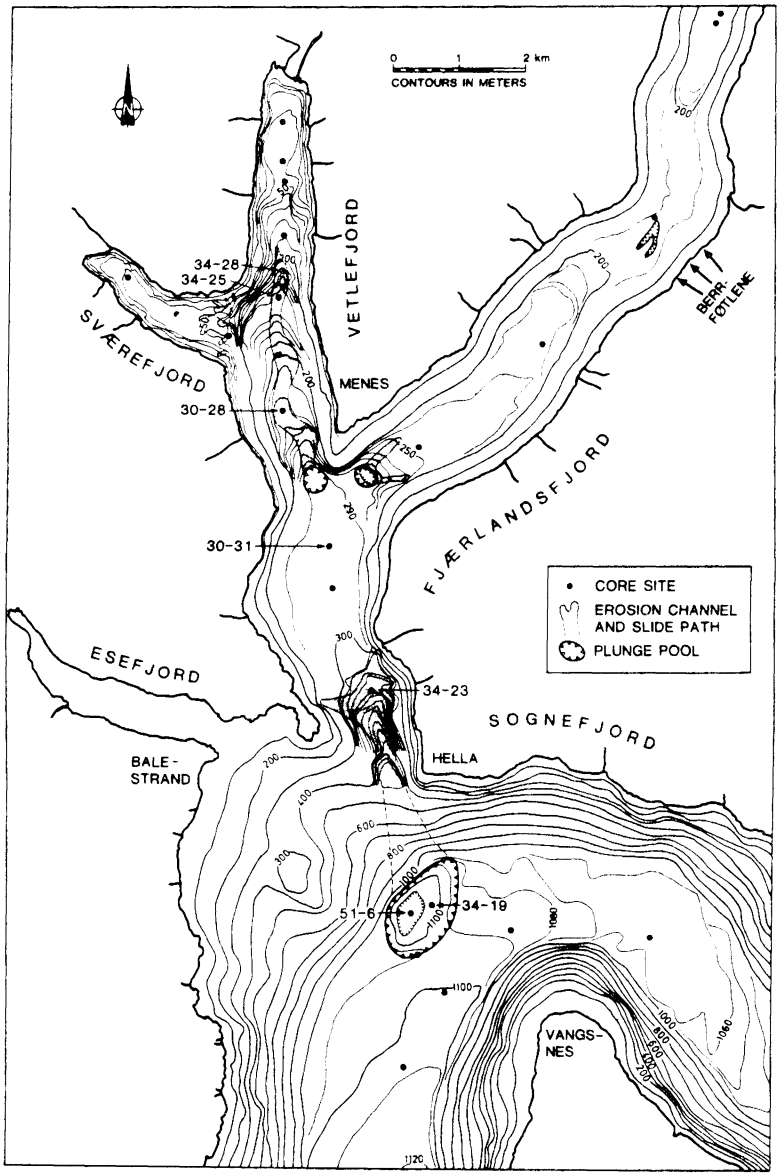


Fig. 36. Map of the outer Fjærlandsfjord, Vetlefjord and Sværefjord showing slide scars and plunge pools from submarine slides. (Modified from Lønne 1981.)

bottomset beds. The sediments are acoustically laminated and sediment cores taken from the southernmost sediment slope also show fine lamination with frequent fine sand laminae in a clayey-silty matrix. The lithology of this part of the sediment sequence indicate a very high sedimentation rate with a rhythmic sediment input from the glacial meltwater channels.

The upper sequence (10-20 m) is characterized by an acoustically homogeneous signature and samples show disturbed sediments with a

mixture of glaciomarine and postglacial sediments (Fig. 37). The upper right part of this profile, (from Vetlefjord) shows clear evidence of extensive sliding activity, as indicated on the map (Fig. 36).

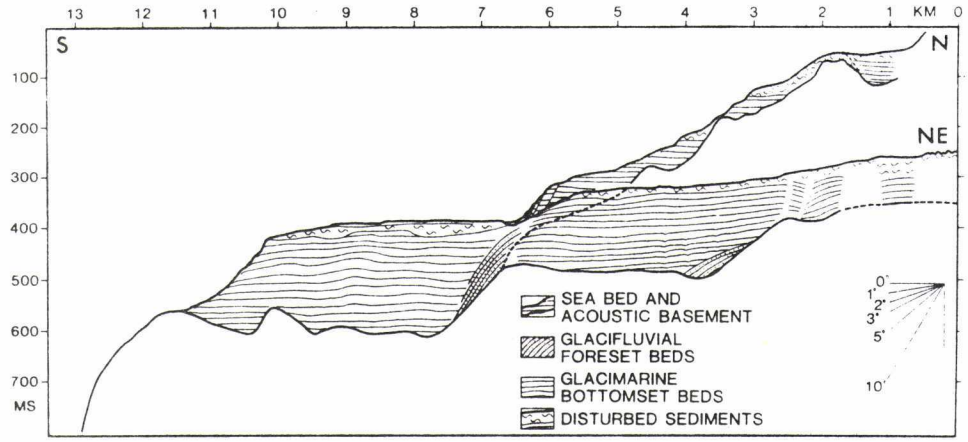


Fig. 37. Interpretation of seismic profiles along outer Fjærlandsfjord and Vetlefjord (upper right). (Lønne 1981.)

The greatest slides have taken place in the 800 m high slope down to the bottom of the Sognefjord proper. A total of 30 x $10^6 m^3$ has slid down the approximately 25^0 steep bedrock slope to the sediment filled fjord channel. Here, a 1 km^2 "plunge pool" was formed as the material struck the soft sediment floor (Figs. 36 and 38). The morphology of this depression indicates that some of the slide material still rests as a 10 m x 200 m sediment lump in the middle of the impact crater. Outside the mound, sediment cores show multiple turbidite sequences with gravel and cobble sized particles.

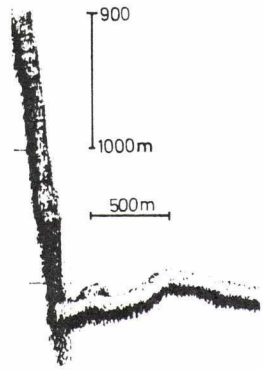


Fig. 38. Echo profile of the "Plunge pool" in soft sediments on the bottom of the Sognefjord, caused by a giant slide 800 m down from the Fjærlandsfjord.

The reason for the sliding is thought to be due to earthquakes causing liquefaction of the fine sand laminae near the sediment slope. The pore pressure in these rather porous laminae is high, due to low permeability in the rest of the sediments and heavy overburden. Thus the adhesive silty clay could move at a very low angle out to the steeper sediment slope. Similar slide scars and plunge pools are recognized in several of the fjord junctions along the Sognefjord where the tributary fjord basins are filled up with sediments.

Sedimentation rates in the Sognefjord

Dateable materials are scarce in the fjord sediments. At 1000 m water depth outside Hermansverk a sediment core contained a moss layer 2 m down in the core at the top of a turbidite sequence. This material must have been deposited as a result of a terrestrial flood from the river Henjaelvi from the north. C^{14} dating of these bryophytes gave 6 250±210 radiocarbon yrs B.P. (Kirkhus 1980) giving a mean rate of sedimentation of 0.3 mm $year^{-1}$ for this period at this locality. Generalization of such values, however, is not possible because of the great variation in the sedimentation processes as described above. Even bottom current erosion is taking places in some areas. But the order of magnitude shows that the present rates of sedimentation at the bottom of the Sognefjord are very low.

A sediment threshold in the Sognefjord near Fresvik

The longitudinal profile of the Sognefjord (see Holtedahl this volume) has a smooth basin character except for a small threshold at 900 m water depth near Fresvik, som 160-170 km from the North Sea. Seismic profiles (Fig. 39), shows that this threshold is mainly made up of fine sediments of the same character as in the rest of the fjord. The acoustic layer configuration does not correspond to a normal basin sedimentation as the nearly horizontal layers end abruptly on the slope down to the basin inside the threshold. There are no signs of sliding along this slope. A standstill of the glacier front in the main fjord could produce

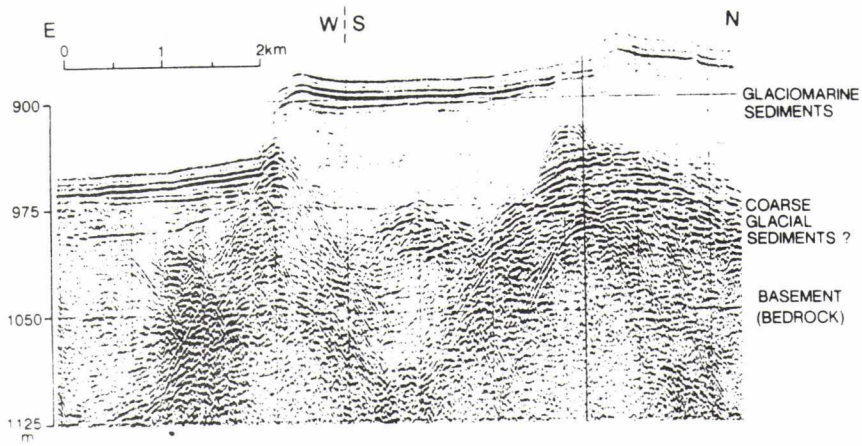


Fig. 39. Seismic profile along the bottom of the Sognefjord near Fresvik showing a threshold of glacial sediments possibly caused by a short stagnation in the glacial retreat about 10000 years ago.

such layering in front of the grounding line of the fjord glacier.

Bergstrøm (1975) mapped the course of the deglaciation in the district of Aurland where he found the highest lateral moraines at 1250-1300 m a.s.l. He suggested that they correspond to a short stagnation of the glacial retreat at the end of the Younger Dryas Chronozone, with the glacier terminus laying at the entrance of the Aurlandsfjord, only a few km from the described sediment threshold. The question of glaciological interest is whether a calving glacier front in a 1100 m deep fjord could reduce its retreat or even make a short stop. The sediments on the bottom of the fjord, as well as the lateral moraines in Aurland, are strong evidence in favor of this.

FLÅMSDAL AREA

Flåmsdal is a typical example of a West-Norwegian fjord valley, deeply cut back into the old "paleic" surface. It extends in a southerly direction from the inner end of Aurlandsfjord (a tributary to the Sognefjord) and terminates in two preciptious valley heads, more than 300 m high, close to Myrdal railway station. The contrast between the broad, smooth Moldå valley of the mountain plateau and its continuation in the Flåm valley is well seen at Vatnahalsen station. The old valley generation to which the Moldå valley belongs, is seen as ledges about 800 m a.s.l. on the valley sides of the Flåm valley. The formation of the steep back wall of this valley, as well as other similar valleys of western Norway, has been much discussed among geomorhologists. It has been explained as the result of headward corrie-erosion, erosion by the inland ice sheet, a combination of fluvial erosion and glacial action (especially by glacial meltwater streaming down in cracks in the ice), and primarily by fluvial erosion. With regard to the inner end of the Flåm valley, its great width and large dimensions strongly point to a glacial origin. The valley itself is also a typical glacial valley, with steep valley sides, a broad valley floor, and an uneven longitudinal profile with steps and riegels (rock bars) cut through by the river in narrow gorges.

It is obvious, however, that fluvial erosion has been of considerable importance in the formation of the valley: this is seen particularly well on the valley floor from Berekvam station and northwards, where Cambro-Silurian phyllites exist. Here the river has cut narrow, deep gorges, in some cases only about one metre wide, and with vertical or even overhanging walls with remnants of potholes. The present path of the river at Berekvam (loc. 12), in a narrow canyon cut down in a ledge which is above the lowest part of the valley floor, suggests it had its origin in a sub-glacial meltwater stream. This is also suggested by remnants of old channels, often with only one wall, which show signs of pothole erosion. The channels run both along the valley and more or less perpendicular to it. The outcropping phyllites, often with fairly flatlying layers, and major fractures in a direction across

the valley, show glacial sculptured surfaces and a conspicious step-like longitudinal profile. The stoss and lee topography points to an ice movement downstream in a northerly direction. The frequently steep valley sides occasionally show marks of fluvial erosion with waterworn surfaces and remnants of giant potholes. These potholes were probably formed by sub-glacial meltwater streams cascading down in channels, on the one side supported by the rocky valley side, and on the other by the ice. Glacial striae shows that there was movement of the ice during or after these fluvial erosional forms were made.

BRIGSDALSBREEN

Introduction

Brigsdalsbreen is located in the upper part of the scenic valley Oldedalen on the western side of Jostedalsbreen. This outlet glacier flows from the highest elevation 1910 m on the ice plateau down to lake Brigsdalsvatnet (350 m a.s.l.) where the terminus is calving (Fig. 40). Below elevation 1500 m the surface is steep and heavily crevassed. In 1966 the glacier covered an area of 10.2 km^2 and had a length of 5.4 km along the central flow line (Pedersen, 1976). The equilibrium line altitude is calculated to 1630 m.

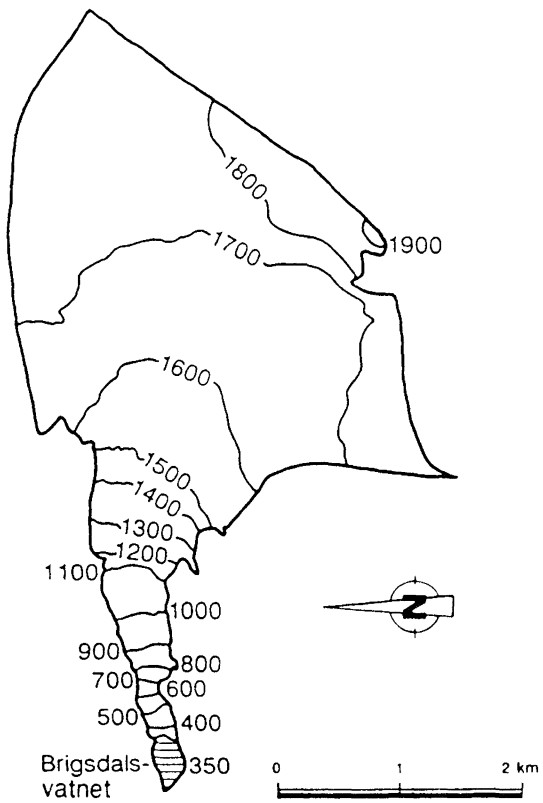


Fig. 40. Map of Brigdalsbreen (after Pedersen, 1976).

Length fluctuations

Annual observations of the front position of Brigdalsbreen started in 1901 (Liestøl, 1961) and are still going on. The cumulative length fluctuation is shown in Fig. 41. The rapid retreat during the period 1930-50 was not only caused by warmer climate, but it was also influenced by calving in Brigdalsvatnet. The maximum depth in the lake has been measured to 27 m. From 1901 the glacier front has receded 800 m. The moraines from several glacier advances are dated from historic documents, dendrochronology and lichenmetry studies and from model calculation of the front position.

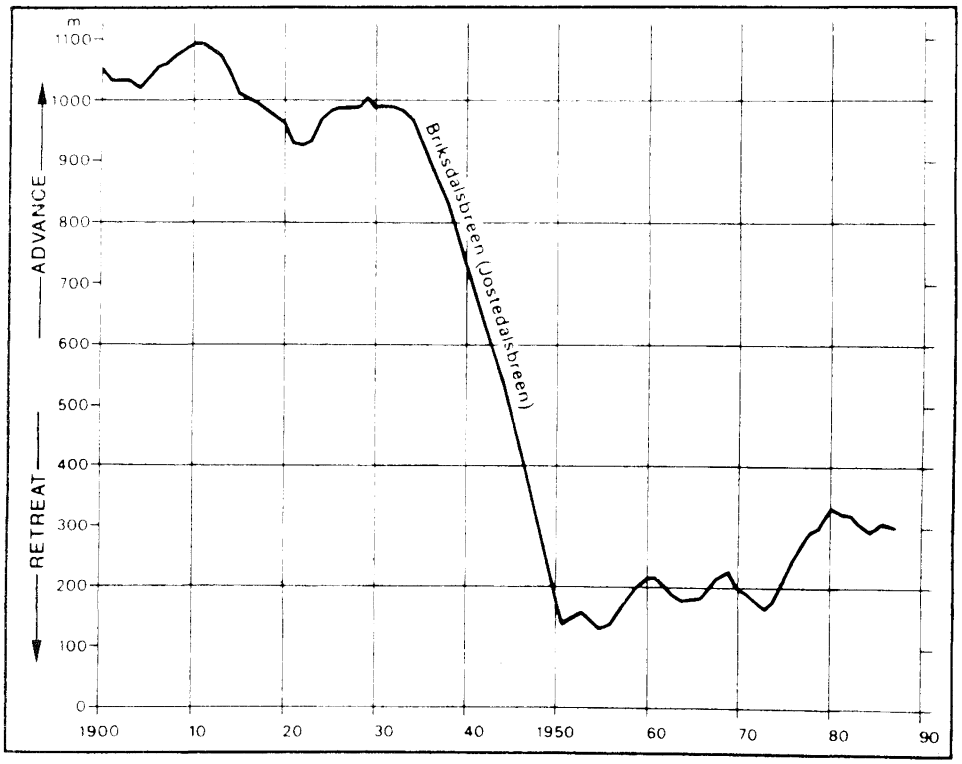


Fig. 41. Cumulative length fluctuation (after Liestøl).

Glacier volume fluctuations

Rogstad (1941, 1951) calculated the annual net balances of Brigsdalsbreen using data from discharge measurements in Oldevatn and Lovatn compared with similar observations in an adjacent glacierfree catchment. Pedersen (1976) carried out mass balance measurements on the upper accessible part of Brigsdalsbreen, which

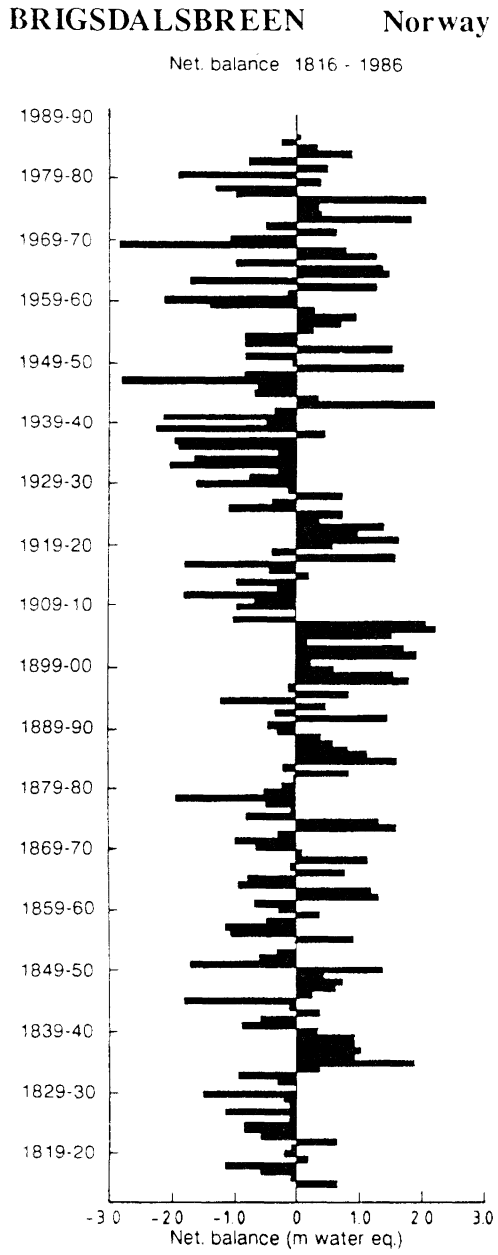


Fig. 42. Calculated annual net balance from 1815-16 to 1986-87 (Based upon Pedersen, 1976 and later computations).

were compared to similar observations on Nigardsbreen (48 km^2). She also recalculated the yearly fluctuation in glacier volume from 1901-02 to 1973-74 based up the "hydrological method" described above. Pedersen found from crosscorrelating annual net balance and length fluctuation that Brigsdalsbreen had a "time-delay" of three years from a change in glacier volume was observed as an advance or retreat at the terminus. Using multiple regression analysis she also found good correlation between annual net balances calculated from winter precipitation and summer temperature from meteorological stations and the change in glacier volume. Thus she calculated the variations from 1816-1974, and the series have then been completed with data from 1975-1986 (Fig. 42). From 1815 to 1834 the cumulative mass deficit on Brigdalsbreen is calculated to about 7 m water equivalent before a positive trend in net balance occurred. The glacier had its maximum volume in 1907 after having its mass increased corresponding to a water layer of 21 m since 1834. A new maxima occurred in 1925 before a warmer climate caused a large mass loss. After 1960 there have been minor oscillations in glacier volume.

REFERENCES

Aarseth,I.
1980: Glaciomarine sedimentation in a fjord environment: Example from Sognefjord. Pp. 16-21 in: Orheim, O. (ed.). Glaciation and deglaciation in Central Norway. Field guide to excursion, Symposium on processes of glacier erosion and sedimentation, Geilo, Norway 1980. Norsk Polarinstitutt, Oslo 1980.

Aarseth,I. and Mangerud, J.
1974: Younger Dryas end moraines between Hardangerfjorden and Sognefjorden, Western Norway. Boreas 3: 3-22.

Ahlmann, H. W:son
1919: Geomorphological studies in Norway. Geogr. Ann. 1 : 1-210.

Bergstrøm, B.
1975: The course of deglaciation in Aurlandsdalen and surrounding areas, West Norway. Norges geol. Unders. 317, 33-69.

Bogen, J.
1983: Morphology and sedimentology of deltas in fjord and fjord valley lakes. Sedimentary Geology 36, 245-267.

Bogen, J. and Olsen, H.C.
1986: Sedimenttransport i driftsvannet til Jostedal kraftverk. Oppdragsrapport 10-86. Hydrologisk avdeling, Norges Vassdrags- og Energiverk.

Bogen, J., Wold, B. and Østrem, G.
1988: Historic glacier variations in Scandinavia. Proc. Symposium on glacier variations and climate. Amsterdam, 1-5 June 1987, D. Reidl, Dordrecht (in press).

Fareth, O.W.
1970: Brerandstadier i midtre og indre Nordfjord. Thesis, Univ. of Bergen. (Unpublished.)

Fareth, O.W.
1987: Glacial geology of the middle and inner Nordfjord. Norges geologiske undersøkelse, Bulletin 408, 55 pp.

Fægri, K.
1933: Über die Längenvariationen einiger Gletscher des Jostedalsbre und die dadurch bedingten Pflanzensukzessionen. Bergen Museums Årbok 1933, Naturvidenskapelig rekke 7, 255 pp.

Grove, J.M. and Battagel, A.
1983: Tax records from western Norway, as an index of little ice age environmental and economic deterioration. Climatic Change 5, 265-282.

Holtedahl, H.
1967: Notes on the formation of fjords and fjord valleys. Geogr. Ann. Ser. A 49: 188-203.

Holtedahl, H.
1975: The geology of the Hardangerfjord, West Norway. Norges geol. Unders. 323. 1-87.

Kirkhus,Ø.
1980: Palynologi i seks sedimentkjerner fra Sognefjorden og kontinetalsokkelen utenfor Møre. Thesis, Univ. Bergen. (Unpublished.) 140 pp.

Krigstrøm A.
1962: Geomorphological studies of sandur plains and their braided rivers in Iceland. Geogr. Ann. 44, 328-346.

Larsen, E., Eide, F., Longva, O. & Mangerud, J.
1984: Allerød-Younger Dryas climate inferences from cirque glaciers and vegetational development in the Nordfjord area, western Norway. Arctic and Alpine Research 16, 137-160.

Lien, R. & Rye, N.
1988: Formation of saw-toothed moraines in front of the Bødalsbreen glacier, western Norway. Norsk Geologisk Tidsskrift 68, 21-30.

Liestøl, O.
1956: Glacier dammed lakes in Norway. Norsk Geografisk Tidsskrift 15, 122-149.

Liestøl, O.
1961: Bremåling og brevariasjoner. Den Norske Turistforenings Årbok 1961, p. 24-43.

Liestøl, O.
1967: Storbreen Glacier in Jotunheimen, Norway. Norsk Polarinstitutt Skrifter 141, 63 pp.

Liestøl, O.
1972: Avalanche plunge-pool effect. Norsk Polarinstitutt Årbok 1972, 179-181.

Lønne, Ø.,
1981: Maringeologiske undersøkelser i Ytre Fjærlandsfjord, Vetlefjord og Sværefjord, Sogn. Thesis, Univ. Bergen. (Unpublished.) 226 pp.

Maizels, J.K.
1979: Proglacial aggradation and changes in braided channel patterns during a period of glacier advance: An alpine example. Geogr. Ann. 61 A, 87-101.

Mundal, E.
1953: Kvartære akkumulasjoner og strandlinjer ved Fjærlandsfjorden og nokre andre lokaliteter i Sogn. Thesis, Univ. Oslo. (Unpublished.)

Nansen, F.
1921: The strandflat and isostasy. Skr. Vid. Selsk. Krist. Mat. - Naturv. Kl 11 :1-313.

Nesje, A.
1984: Kvartærgeologiske undersøkingar i Erdalen, Stryn, Sogn og Fjordane. Thesis, Univ. of Bergen. (Unpublished.)

Nesje, A. Anda, E., Rye, N., Lien, R., Hole, P.A. & Blikra, L.H.
1987: The vertical extent of the Late Weichselian ice sheet in the Nordfjord-Møre area, western Norway. Norsk Geologisk Tidsskrift 67, 125-141.

Orheim, O.
1970: Glaciological investigations of Store Supphellebre, West-Norway. Norsk Polarinstitutt Skrifter 151, 48 pp.

Pedersen,K.
1976: Glasiologiske og glasialgeologiske undersøkelser. Thesis, Univ. i Bergen. (Unpublished.) 246 p.

Raubakken, V. Suspensjonstransport og sedimentasjon i
1985: Erdalsvassdraget, Stryn. Thesis, Univ. of Oslo. (Unpublished.) 106 p.

Rogstad, O.
1941: Jostedalsbreens tilbakegang. Norsk Geogr. Tidsskrift. Bind 8, p 273-298.

Rogstad, O.
1951: Variation in the glacier mass of Jostedalsbreen. Journal of Glaciology, Vol. 1, p. 551-556.

Roland, E.
1986: Antatte endringer på Austdalsbreen og Sygneskardbreen etter regulering av Austdalsvatn og Kupvatn. Oppdragsrapport 13-86. Hydrologisk avdeling, Norges Vassdrags- og Energiverk, 25 + 5 p.

Roland, E. and Haakensen, N.
1985: Glasiologiske undersøkelser i Norge 1982. Rapport nr. 1-85. Hydrologisk avdeling, Norges Vassdrags- og Elektrisitetsvesen, 102 p.

Rye, N., Nesje, A., Lien, R. & Anda, E.
1987: The Late Weichselian ice sheet in the Nordfjord-Sunnmøre area and deglaciation chronology for Nordfjord, western Norway. Norsk geografisk Tidsskrift 41. 23-43.

Röthlisberger, H.
1987: The Jostedal Project. Evaluation of the Need for Further Glaciological Investigations Related to Austdalsbreen and the Styggevatnet Hydropower Reservoir. Rept. No. 98, 4. Versuchanstalt für Wasserbau, Hydrologie und Glaziogie an der Eidgenossische Techn. Hochschule Zürich.

Schieldrop, B.
1986: Austdalsbreen - Bølger som følge av kalving i Austdalsvatn. Appendix in Roland, E., 1986 (see above), 5 p.

Steen, J.-E., and Mathiesen, M.
1987: Kalving fra Austdalsbreen. Måling av bølger, strøm og temperatur. SINTEF-rapport STF60 A87137. Norsk Hydroteknisk laboratorium. (English Summary), 29 + 43 p.

Sætrang, A.
1988: Istykkelsesmålinger med breradar på Austdalsbreen. Oppdragsrapport nr 5-88. Heli-Anlegg A/S. (English Summary), 44 p.

Sætrang,A. and Holmquist, E.
1987: Kartlegging av istykkelse på Nordre Jostedalsbreen. Oppdragsrapport nr. 8-87. Hydrologisk avdeling, Norges Vassdrags- og Energiverk, 15 p.

Vangsnes, T.
1981: Maringeologiske undersøkelser i Indre Fjærlandsfjord, Sogn. Thesis, Univ. Bergen. (Unpublished.) 201 pp.

Vorren, T,
1973: Glacial geology of the area between Jostedalsbreen and Jotunheimen, South Norway. Norges geol. Unders. 291, 1-46.