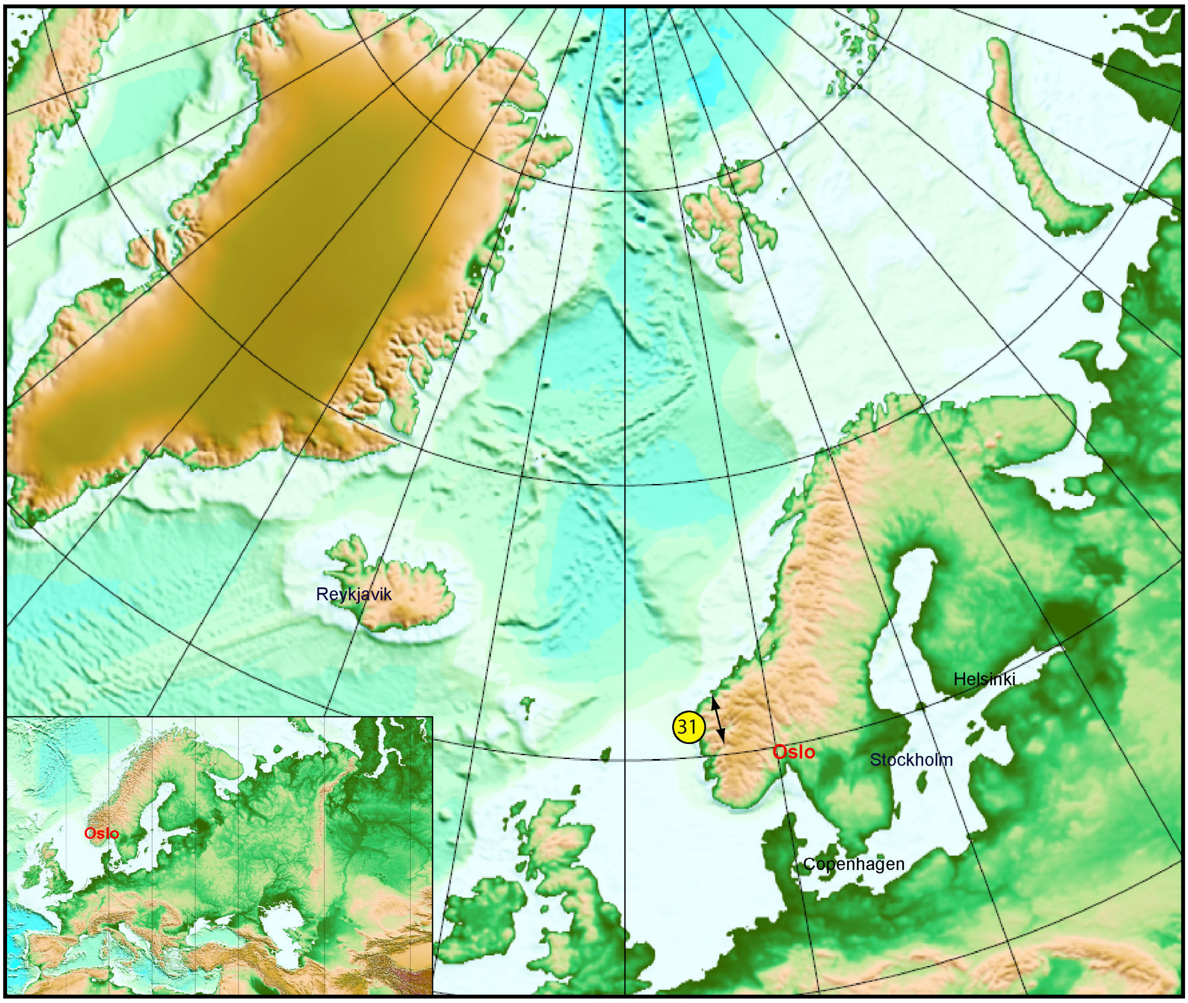
33 IGC excursion No 31, August 14 – 20, 2008

33 IGC, The Nordic Countries

UNESCO FJORDS

From Nærøyfjord to Geirangerfjord

Organizers:
Inge Aarseth, **University of Bergen*****, Atle Nesje*****, University of Bergen and Bjerknes Research Centre, Bergen, Norway** ***and Ola Fredin,*** **Geological Survey of Norway, Trondheim**



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Abstract

In addition to magnificent scenery, fjords may display a wide variety of geological subjects such as bedrock geology, geomorphology, glacial geology, glaciology and sedimentology. This excursion will bring us to the heartland of the western Norwegian fjords including the two fjords listed on the UNESCO heritage list: the Nærøyfjord and the Geirangerfjord. These fjords were selected as the most spectacular among many others because of their natural beauty as well as their cultural heritage. Because of the great relief, both terrestrial and submarine, they are subject to dramatic exogenic processes. The excursion will give us an overview and a detailed insight in the formation of fjords in addition to the geological processes taking place in the fjord environment today. We will see how glaciers respond to climatic change and how man is preparing for natural disasters from snow avalanches as well as from tsunamis caused by rockslides. We will see fjords from below (ship decks), from above (outlook points), as well as from hotel restaurants and you will learn that the saying: "If you have seen one, you have seen them all" does not apply to these fjords.

Logistics

Dates and location

Timing:	**Thursday August 14th – Wednesday August 20th 2008.**
Start location:	**The excursion starts by bus from Bergen Airport Flesland.**
End location:	**The excursion ends outside the Oslo Central Station, Wednesday August 20th before 17:00. There will be a bus stop at Oslo Airport Gardermoen at 16:00.**

Travel arrangements

The participants will be picked up at Bergen Airport Flesland Thursday August 14th immediately after the 8:10 arrival of the morning flights from Oslo. The participants have to book their own flights. (Both SAS and Norwegian depart Oslo Airport Gardermoen at 07:15).

The participants will be brought back to Oslo by bus Wednesday August 20th before 17:00. The bus will stop at Oslo Airport at 16:00 for participants leaving the same evening. Final stop will be outside the Oslo Central Station.

Accommodation

In total six overnight stays are included: At two locations we will stay for two nights. (Nordfjordeid and Stryn). Breakfast and lunch packages will be provided together with the overnight accommodation. On the first day we will have lunch onboard the chartered boat on the Nærøyfjord and the Aurlandfjord. Since this is long after your breakfast, you will have a snack at the first geology stop just north of Bergen. On two of the days we will have lunch served at a restaurant in front of a glacier (included). Breakfast and evening meals (usually a buffet dinner) will be served at hotel restaurants along the route.

Since there will be no cultural stop in the city of Bergen, there is a possibility to visit Bergen on the 13th ("The capitol of the fjords"), by leaving Oslo a day in advance and join the excursion in Bergen. Pick up point is on the west side of (outside) Bergen Bus Terminal at AM 09:00 on August 14th. (The street's name: "Fjøsangerveien"). Participants not arriving Bergen by air Thursday morning must contact inge.aarseth@geo.uib.no or mob. phone: +47 4684 3311.

Field logistics

The excursion area is accessible by bus (Fig. 1). In addition there will be two 2-hours boat trips (on the Nærøyfjord and the Geirangerfjord). In addition three fjord crossings by ferry. The programme in the excursion area will include little hiking except for two days: Saturday August 16th at Vågsøy island on the coast (2 hours) and Wednesday August 19th in easy terrain in Erdalen valley (5-6 hours). Some of the public roads are typical narrow western Norwegian roads with hairpin bends up or down steep mountainsides, but the 12.4 m long bus has licence on these roads. For personal equipment one should bring mountain boots, warm clothing and rain gear. We will visit mountains above 1400 m a.s.l. and it might be cold up there even in mid August.

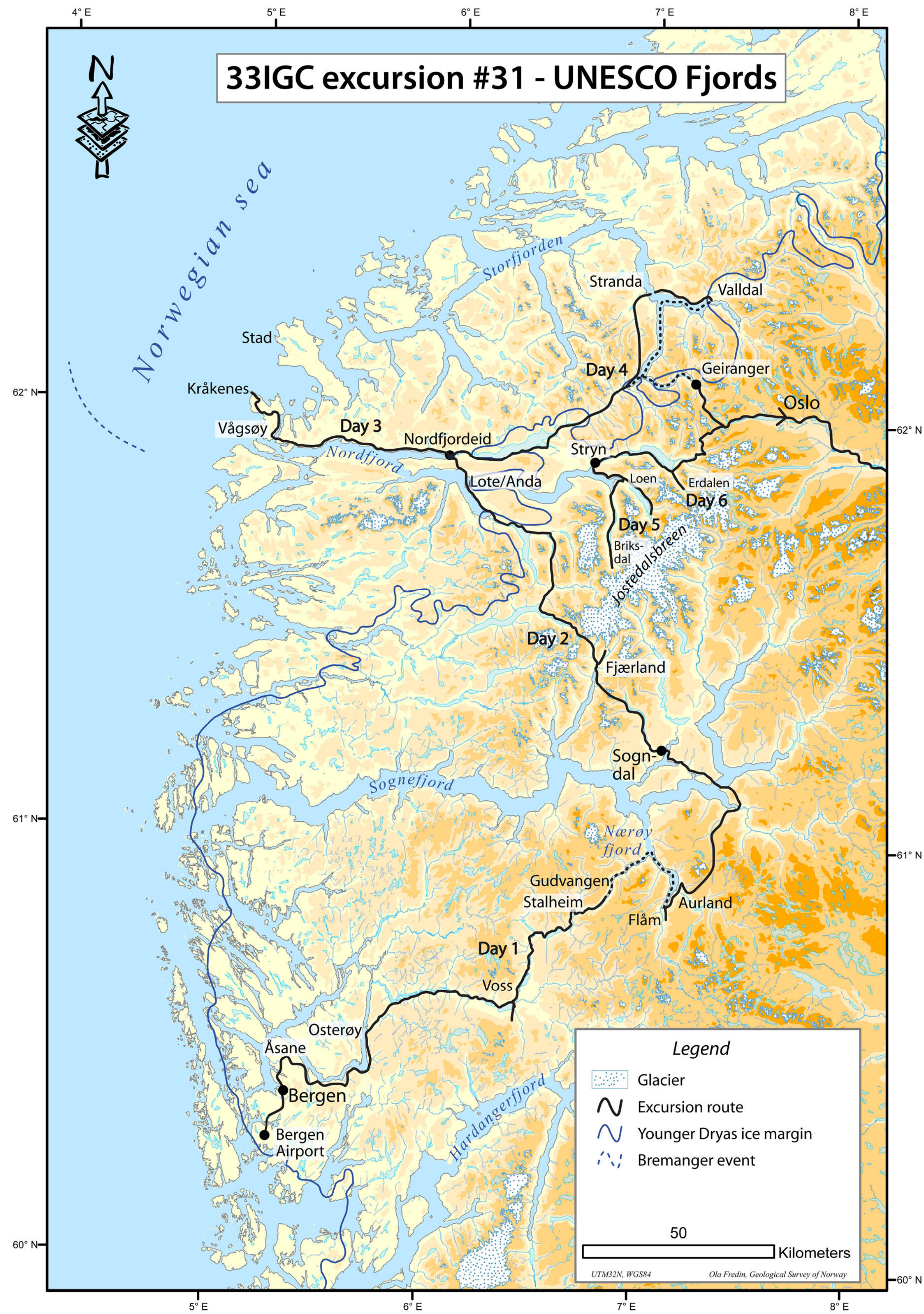

Fig. 1. Excursion route. Dotted: Fjord cruises. Glacier margins of the Younger Dryas stadial and the Bremanger event are indicated.

General Introduction

This excursion will take you to the two fjords listed on UNESCO heritage sites: The Nærøyfjord and the Geirangerfjord as well as the fjord areas in between. We will also visit the coast at Vågsøy island (62 ºN) just south of Cape Stad. The excursion will provide a detailed and integrated overview of processes, deposits, landforms, human impact and landscape development in the heartland of the fjords. It will include the discussion of long-term pre-Quaternary landscape development, Quaternary glaciations, deglaciation as well as present-day surface processes including denudation and accumulation rates. The area has long been the focus of Quaternary, marine as well as geomorphologic studies. Ongoing research programs deal with several topics such as glaciers reaction to global warming, geohazards from snow as well as from rock avalanches (historic and potential tsunamis) and chemical and physical weathering.

Regional Geology

Since the excursion deals with several topics, the regional geology will be split up in several chapters written by geologists working on the subject in the area.

The bedrock of the Bergen-Geiranger area

by Haakon Fossen[1]

The bedrock reflects the general tectono-stratigraphy of the Scandinavian Caledonides; a Precambrian basement (Western Gneiss Region: WGR) that were increasingly involved in Caledonian deformation and metamorphism to the west, an overlying unit of micaschist and phyllite, and remnants of Caledonian Nappes that were thrust above the micaschist/phyllite layer and the basement (Fig. 2).

The *Western Gneiss Region* consists of intrusive complexes and subordinate metasedimentary rocks that were deformed and to a large extent turned into gneisses during Proterozoic as well as Caledonian orogenic movements. It was once covered by several tens of kilometers of Caledonian allochthonous units (nappes) that constituted the Caledonian orogenic wedge. Today the WGR defines a major window in the Scandinavian Caledonides. Eclogites and high-pressure minerals such as micro-diamonds and coesite in the northwestern part of the WGR suggest that this represents the paleo-margin of Baltica that was subducted underneath Laurentia to depths in excess of 100 km.

A layer of phyllite and micaschists separates the Caledonian nappes from rocks of the WGR. This mechanically weak layer acted as a décollement during Siluro-Devonian Caledonian collision, during which a tectonic wedge of allochthonous units accumulated. The Caledonian allochthons (nappes) constitute fragments of the Baltica margin and oceanic crust from the lower Paleozoic Iapetus ocean, including island arc complexes as well as ophiolite fragments. The largest nappe unit is the Jotun Nappe, which occupies the eastern Sogn area. [1]

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[1] University of Bergen

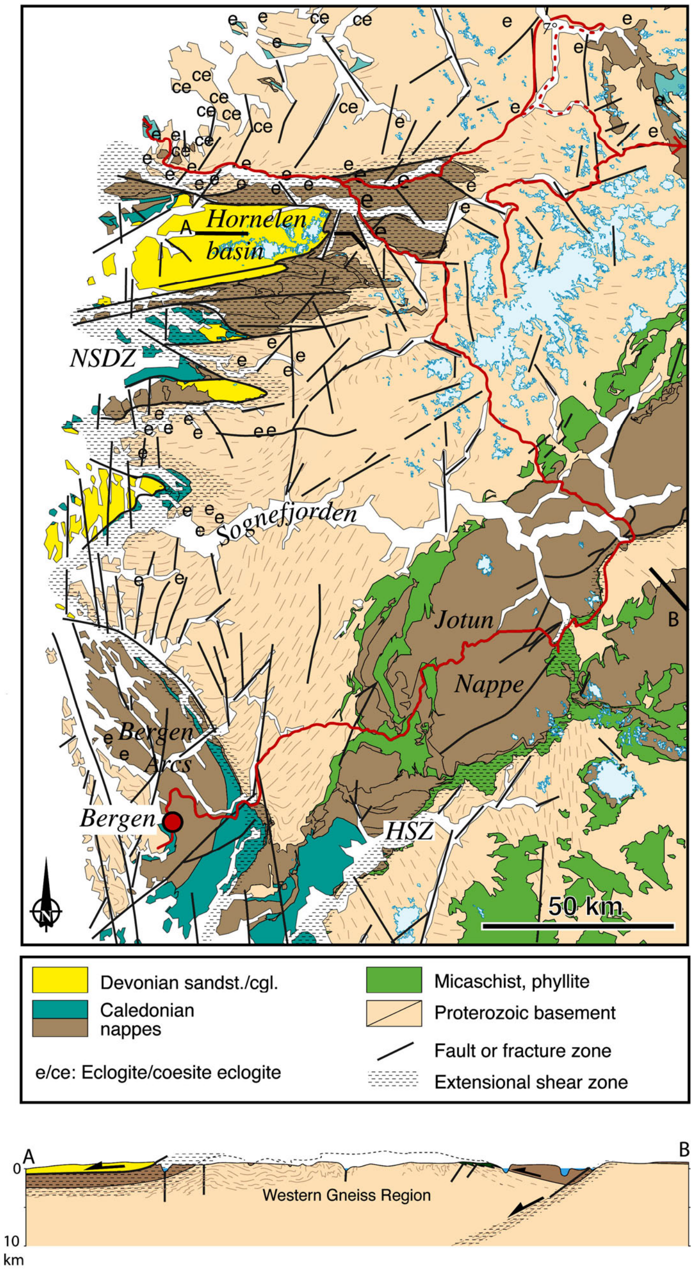

Fig. 2. Bedrock geology of the excursion area. Below: W-E-SE profile of the bedrock units. HSZ: Hardangerfjord Shear Zone.

Shear zones, faults and fractures

At the end of the Caledonian orogeny (early Devonian times), Caledonian contraction gave way for Devonian extension with the formation of impressive extensional shear zones. The initial stage was characterized by massive back-sliding of the orogenic wedge on the micaschists/phyllites (décollement zone). Then W- and NW-dipping extensional shear zones evolved, notably the Hardangerfjord Shear Zone (HSZ) and the Nordfjord Sogn Detachment Zone (NSDZ). The former is seen in the Aurland-Lærdal area, where the basement-phyllite interface bends down towards the Sognefjord. The NSDZ involves many tens of kilometers of offset and separates Devonian conglomerates and sandstones in its hanging wall from (ultra) high rocks, notably eclogites, in the WGR. The HSZ involves a few kilometers of offset, and marks the transition from the variously deformed and rotated WGR to almost unaffected basement to the SE.

Reactivation of these shear zones as brittle faults occurred repeatedly after the Caledonian orogeny. In addition, many post-Devonian faults and fractures, particularly along the coast, have coast-parallel trends and have been related to the North Sea Permo-Triassic and late Jurassic rifting events. The many structural “grains” that have resulted from the prolonged structural history of this area are reflected by the different orientations of valleys and fjords, including the many “kinks” on the Sognefjord.

Geomorphology

by Inge Aarseth

Geomorphologic studies in the fjord region of western Norway have been undertaken by several authors. The older works took very little account of the bedrock geology. Ahlman (1919) was in favour of relative modest glacial erosion except for the over-deepening of the fjords. Gjessing (1967) on the contrary pointed to glacial erosion as the most active process in landscape development along fjords and valleys. Valley heads were according to him formed by backward glacial erosion (Fig. 3). The mountain plateaus, like the Hardangervidda plateau, had not been altered much since early Cenozoic time except for deep weathering. The weathered material had been removed later by periglacial mass movement as well as glacial erosion (Gjessing 1967). Holtedahl (1967, 1975), who studied the area around the Hardangerfjord in particular, found several areas of potholes and other P-forms and argued that glaciofluvial erosion had played an important role in fjord formation. According to him confluence of tributary glaciers in the inner fjords led to excavation of deep fjord basins. Near the coast the glaciers spread out and lost their concentration. This diffluence reduced the erosion and left behind bedrock thresholds. Nesje & Whillans (1994) argued that V-shaped valleys and gullies along the sides of the Sognefjord pointed to a strong fluvial and downslope action. The glaciers were only responsible for removal of the weathered material and the over-deepening of the fjords. They connected the gently sloping surfaces above the fjord- and valley shoulders and constructed the preglacial landscape, the paleic surface. A resent referee paper gives a comprehensive discussion on the Atlantic coast and fjords (Corner 2005a).

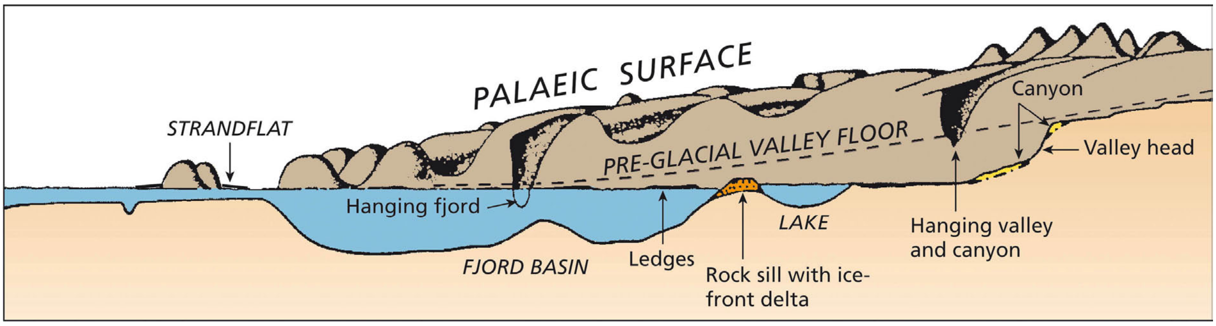

Fig. 3. Schematic profile of a fjord showing the paleic surface and the Norwegian strandflat. (Modified from Gjessing (1967).

When comparing a satellite image of the fjord district (Fig. 4), to a bedrock map of the same area (Fig. 2) one realize the connection between the bedrock structures and the architecture of the fjords. In recent years several authors have linked the upper terrestrial bedrock surface (the paleic surface) to the history of the offshore sedimentary basins (Doré 1992). During the Late Cretaceous transgression most of Southern Norway was covered by Cretaceous sediments. This was found by extrapolating seismic reflectors, like the Cretaceous/Tertiary

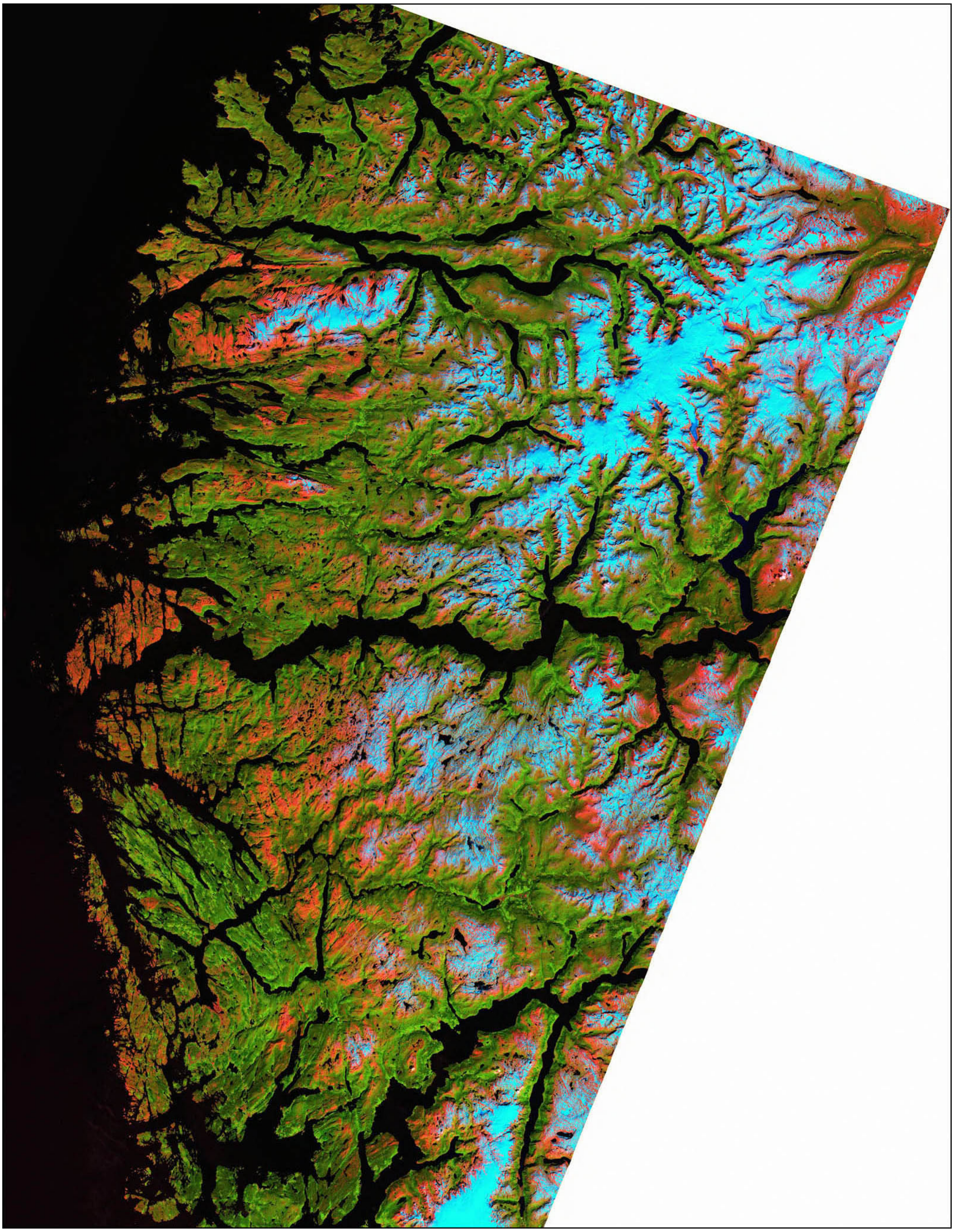

Fig. 4. Landsat 7 false colour mosaic (picture date 2007-07-21) showing south-western Norway and most of the excursion area. The colours (Landsat bands 7,4,2) essentially show snow and ice as cyan, bedrock and soil as red and vegetation as green. Water is black. Note the structurally controlled fjords e.g. around the city of Bergen and the "dendritic" appearance of the Sognefjord.

boundary from the North Sea on to the land area. These sediments were stripped off during the following Cenozoic uplift caused by the opening of the Norwegian-Greenland Sea. The uplift has been tectonically pulsating. A convoluting surface of the terrestrial mountains is thought to represent the uplifted Early Tertiary surface (Doré 1992). The surface was exposed to weathering, mass wasting and fluvial transport to the sedimentary basins in the North Sea. The transport and deposition of sediments caused down-warping of the basin and uplift of the terrestrial surface. During the glaciations this process increased when the deep valleys and fjords were formed. Calculations of sediment wedges offshore have led to estimates of a mean onshore erosion of 1000 m in the area just inside the coast of western Norway (Riis 1992). A resent review paper gives a comprehensive discussion on the geomorphology of the Scandinavian mountain areas (Corner 2005b).

The Norwegian strandflat

by Inge Aarseth

Excursion # 31 starts at the Bergen Airport, Flesland. The airport is situated on the Norwegian strandflat, a term introduced by Reusch (1894). This rock terrace extends along the coast of Norway from just north of Stavanger all the way to the North Cape, but not further east. Generally, the strandflat is situated at 30-60 m a.s.l., but in some areas it is partly submerged with shoals and deeper submarine bedrock terrain down to about 40 m water depth. The width also varies considerably from a narrow rim near Vågsøy island (the third day of the excursion), to 20-40 km across just north of Bergen, along the coast of Trøndelag (63 – 64 ºN) and at Helgeland just south of the Polar Circle. The early debate on the formation of the strandflat was based on sparse knowledge of geological stratigraphy and climate history, and a wide range of processes as well as time frames were suggested for its formation. Nansen (1904, 1922) was in favour of a relatively young (Quaternary) age. His main argument was that the strandflat was formed after the fjords and sounds between the islands. This dissection of the coastal landmass had also made the actual extent of the strandflat possible (Fig. 5). He also pointed out the effect of glacial isostasy resulting in a general coastward tilt of the strandflat.

In more recent years sediments dated from marine caves suggests an Early Weichselian (last glaciation) minimum age for the strandflat, and frost weathering along the shore has been considered the main mode of formation in addition to wave abrasion (Larsen & Holtedahl 1985). Average glacial coverage through the Quaternary left the coastline free from glaciers for long periods and this allowed the processes responsible for the formation of the strandflat (Porter 1989). Holtedahl (1998) found a close connection between strandflat heights and Late Weichselian marine limit south of 64 ºN. He did not exclude an older formation of the strandflat in some areas where it nearly coincides with a pre-strandflat paleo-surface, thought to be of pre-Jurassic age.

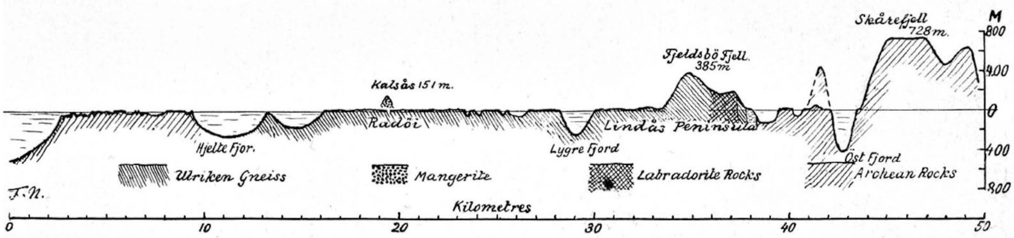

Fig. 5. Profile across the Norwegian strandflat 40 km north of Bergen (Nansen 1922).

The deglaciation of the area between Hardangerfjorden and Storfjorden, western Norway

by Jan Mangerud[2]
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Between Hardangerfjorden and Storfjorden (Fig. 1) the Scandinavian Ice Sheet reached the edge of the continental shelf during the Last Glacial Maximum (Mangerud 2004; Nygard et al. 2004; Ottesen et al. 2005). The ice sheet was on the shelf characterized by fast flowing ice streams separated by areas of slower ice. Most of these ice streams followed troughs that are continuations of the fjords, but in the Norwegian Channel there was an ice stream parallel to the coast. According to Nygard et al. (2004) the ice margin started to withdraw from its maximum position soon after 15 ^{14}C ka (about 18 cal ka) and there was a re-advance (the Bremanger event, Fig. 1) across much of the shelf before 13.3 ^{14}C ka.
Along the coast a number of samples of marine molluscs and terrestrial plant remains have given ages in the range 12.4-12.9 ^{14}C ka, indicating that the coast first became ice free soon after 13 ^{14}C ka (Svendsen & Mangerud 1987; Kristiansen et al. 1988; Mangerud 2000; Nygard et al. 2004; Bondevik et al. 2006), although in the Bergen area the ice margin re-advanced to the open ocean again some time between 12.1 and 12.4 ^{14}C ka (Fig. 6). When the ice margin had withdrawn inside the shallow sills the glacier tongues would have to float in the deep fjords. This lead to a fast calving and in most areas glacial striae show that the fjords calved up first and that the last ice flow was from the adjacent land areas towards the middle of the fjords.

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[2] University of Bergen

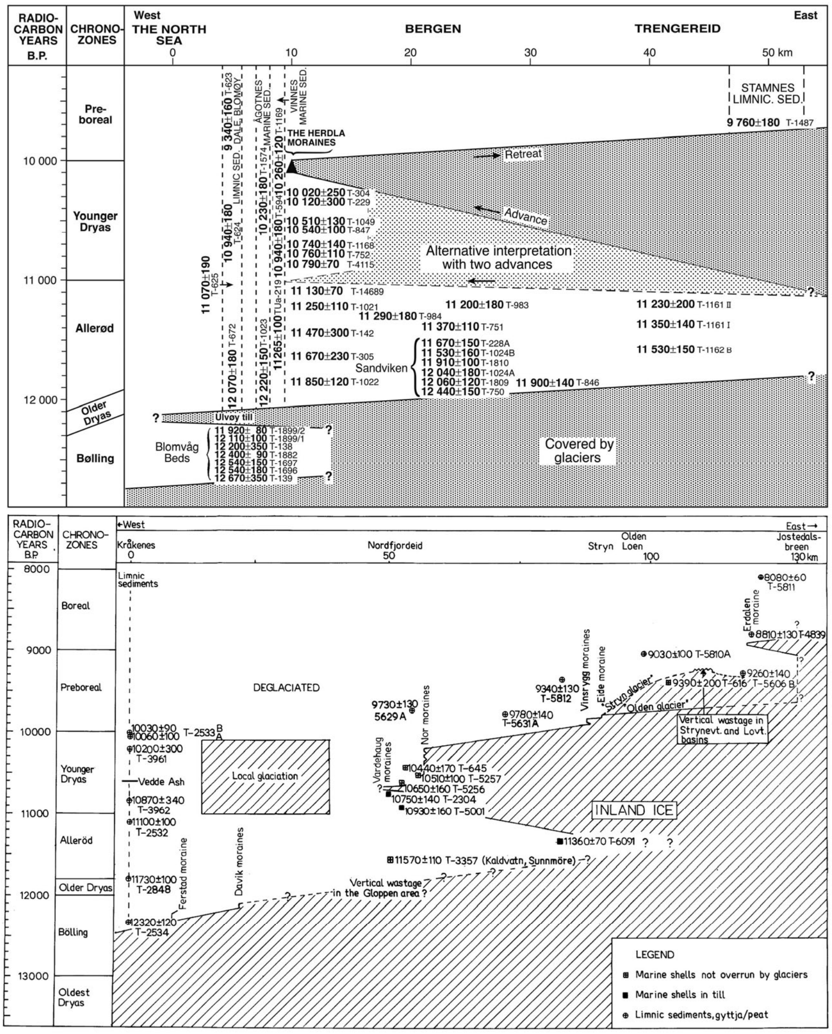

Fig. 6. The upper diagram shows ice-front fluctuations in the Hardangerfjord-Bergen district, drawn in a schematic cross section from the open ocean to the left and into the fjords to the right. Slightly modified from Mangerud (1980). The lower is a similar diagram for the Nordfjord area from Rye et al. (1987).

During the Allerød the ice margin withdraw far into the fjords in the entire area. However, the behaviour of the ice sheet during the Younger Dryas was very different between the southern (Hardangerfjorden-Bergen) and northern (Nordfjord- Storfjorden) areas (Fig. 6). In the southern area the ice margin re-advanced almost to the open ocean and in several fjords that were ice-free during the Allerød, the ice obtained thicknesses of well above 1000 m (Andersen et al. 1995). The re-advance caused a relative sea level rise of 10 m, but the sea-level data also indicate the re-advance in fact started during later parts of the Allerød (Lohne et al. 2007). The ice margin reached the maximum position at the very end of the Younger Dryas (Aarseth & Mangerud 1974; Bondevik & Mangerud 2002). In the northern area, in contrast, the ice margin reached the maximum position early in the Younger Dryas (Fig. 6), the moraines are located far into the fjords (Fig. 1) and there was no sea-level rise (Svendsen & Mangerud 1987). The different reactions of the ice sheet were probably partly due to higher precipitation in the southern area and partly to topographic differences (Mangerud 1980). In the southern area large mountain plateaus close to the coast acted as accumulation areas when the equilibrium line was lowered, whereas the more alpine topography in the northern area mainly generated ice caps and cirque glaciers during the Younger Dryas (Larsen et al. 1998; Sønstegaard et al. 1999).

Glaciers and climate

by Atle Nesje

Glaciers respond to both summer temperature and winter (accumulation-season) precipitation as expressed in the glacier mass balance. In Norwegian mass-balance studies, the ablation and accumulations seasons are 1 May-30 September and 1 October-30 April, respectively (Kjøllmoen 2004). The balance year goes from early October to end of September the following year. The response time or time lag of a glacier is the lag between a climate change and observed length change at the glacier front. This time lag is highly variable, ranging from 3-4 years for steep and short glacier snouts round Jostedalsbreen (Nesje et al. 1995) to 15-60 years on longer, more gently sloping maritime glaciers (Johannesson et al. 1989).

Analyses show that the net mass balance of maritime glaciers in southern Norway is mainly controlled by the winter balance (Nesje et al. 1995, 2000). At Ålfotbreen glacier, located at the coast of western Norway just south of Nordfjord, the correlation coefficient between the winter balance and the net balance and between the summer balance and the net balance (1962-2003) are 0.85 and 0.63, respectively. In the extratropical Northern Hemisphere, two related major weather modes, the North Atlantic Oscillation (NAO) and the Arctic Oscillation (AO), have been related to inter annual (winter-season) temperature and precipitation variability (Hurrell et al. 2003). The NAO is commonly described as the sea-level pressure difference between Iceland and the Azores, and relates to the strength of the westerly winds across the North Atlantic (Hurrell 1995; Luterbacher et al. 2002). In a positive NAO phase, the meridional air-pressure gradient is large in winter in the North Atlantic region, bringing mild and humid winter weather in south-western Scandinavia.

Inter-annual and decadal variations in glacier mass balance in western Norway in the late 20th/early 21st centuries have been attributed to the NAO. Positive NAO index winters yield above normal winter accumulation, and (if not compensated by a following warm summer), positive net mass balance on coastal glaciers in southern Norway, but low winter accumulation on glaciers in the Alps, and *vice versa* (Nesje et al. 2000; Reichert et al. 2001; Six et al. 2001). The correlation coefficient between the annual NAO index (Nesje et al. 2000a) and the winter balance of Ålfotbreen and Nigardsbreen from 1962 to 2003 are 0.77 and 0.71, explaining 59 and 50% of the variability, respectively. In the early 1990s, prevailing positive NAO index winters caused glaciers in western Norway with short response time to advance.

Marine geology

by Inge Aarseth

The fjord sediments of western Norway have been subject to geological investigation for many years (Holtedahl 1967, 1975), from the first use of simple echo sounders to the present day technology with Multi beam echosounder and TOPAS (Parametric Sub-bottom Profiler System) PS18 (Hjelstuen et al. in prep). The fjords act as effective sediment traps during deglaciation as well as during interglacial and interstadial phases. Acoustic and sedimentologic investigations have been used in estimating quantities (Fig. 7) and composition of sediments present in the fjords today (Aarseth 1997). Correlation to raised marine sediments along the fjords as well as chrono- and biostratigraphic analyses of sediment cores have made determinations of ages and sedimentation rates possible (Sejrup et al. 1996).
Sedimentary infill in some coast-parallel fjords is correlated to terrestrial sediments older than Late Weichselian. Approximately 10 % of the fjord sediments in western Norway predate the Weichselian deglaciation (Aarseth 1997). During the deglaciation glaciers were subject to several oscillations (Mangerud 1970) of which the Allerød/Younger Dryas can be traced in terrestrial as well as in fjord sediments. In many fjords the moraines deposited during

Younger Dryas comprise large accumulations with glaciofluvial foreset beds 50-100 m thick resting on even thicker bottomset beds. The distal glaciomarine sediments may amount to 350 m in some fjords in western Norway. These sediments typically consists of 25-55 % clay, 70-45 % silt and < 2-3 % fine sand except for occasional ice-rafted gravels (Aarseth et al. 1989).

High sedimentation rates during the deglaciation created sediment slopes subject to gravity failure. The largest slides took place on slopes connecting tributary fjords and much deeper trunk fjords. Most slide activity happened during and just after the deglaciation, but slides up to 15 x 10^6 m^3 took place a few decades ago in the inner parts of Nordfjord (Aarseth et al. 1989). Fjord sediments have also been used to demonstrate climatic variation using biostratigraphy as well as oxygen isotopes (Mikalsen & Sejrup 2000). Present sedimentation rates in the deep fjords are in the order of 0.3 – 1.0 mm/yr. In fjords fed by rivers draining glaciated areas sedimentation rates may be 1-4 mm/yr (Aarseth 1997). Fig. 8 shows a schematic longitudinal profile of an archetypical western Norwegian fjord (Sejrup et al. 1996).

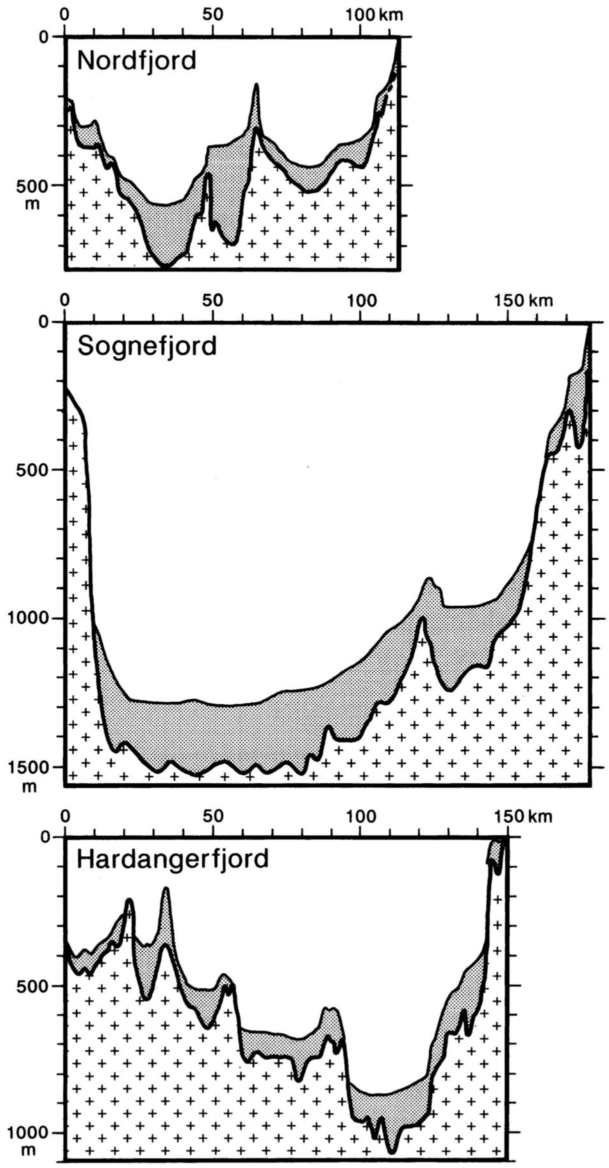

Fig. 7. Sediment distribution in the three longest fjords in western Norway (Aarseth 1997).

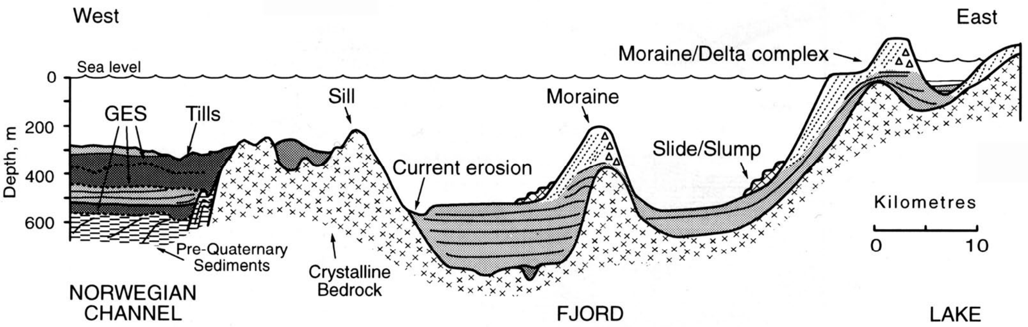

Fig. 8. Schematic longitudinal profile of an archetypical western Norwegian fjord (Sejrup et al. 1996).

Excursion Route and Road Log

The excursion route is marked on Fig. 1. Bergen airport to town centre (20 km). Further along the E-16 (European road 16) from Bergen to Voss and Gudvangen (138 km). Then follows a 30 km fjord cruise on the UNESCO site Nærøyfjord and Aurlandfjord to Flåm, from where the bus takes us 45 km on the “Snowroad” (County road 243, instead of the 24,5 km long Lærdal tunnel – the longest in the world), over 1300 m high mountains to Lærdal and across the fjord to Sogndal. (230 km by bus).

On the second day we will drive on the “Geo nature route” (R-5) to Fjærland. Here the Norwegian glacier Museum will be visited + two outlet glaciers. Then further under the Jostedalsbreen glacier to Skei to pick up E-39 towards the ferry crossing Anda – Lote just before Nordfjordeid. 135 km.

On the third day we will drive towards the coast (The North Sea) on the northern side of Nordfjord (R-15). From the town of Måløy on the island Vågsøy we drive on local roads to the westernmost point Kråkenes. On Vågsøy coastal abrasion, weathering pits and Younger Dryas local glaciation will be studied in detail. Back to Nordfjordeid. 160 km.

The fourth day will take us east on R-15 and R-60 to Hellesylt and Stranda where we will get information on the Åknes – Tafjord rock slide monitoring and early-warning centre. The Geirangerfjord will be studied from a 50 km long afternoon ferry trip from Valldal to Geiranger. (110 km by bus)

The fifth day starts with three breathtaking views of the Geirangerfjord: The “Eagle Road Bend”, Flydalsgjuvet canyon and if weather permits also Dalsnibba (1476 m a.s.l.). Further to the Strynefjell mountain road snow avalanche research station**,** Briksdalsbreen glacier and Loen (historic rock falls and tsunamis). Overnight stop at Stryn. 195 km.

The last day in the field will be spent in the tributary valley Erdalen and at the National Park Centre at Oppstryn on R 15. Back to Stryn. 60 km.

On August 20th the bus will bring you back to Oslo airport (16:00) or the centre of Oslo (17:00). 490 km.

Excursion Stops

Day 1: Thursday, 14th August :
(Oslo) - Bergen Airport – Nærøyfjord – Flåm - Sogndal

Introduction

The first day of excursion # 31 “UNESCO FJORDS: From Nærøyfjord to Geirangerfjord” will take us from Bergen Airport, Flesland, situated on the coastal platform “The Norwegian strandflat” to the Sognefjord with maximum glacial erosion. Here the relief is 2800 m in one slope, from mountain peaks at 1700 m a.s.l. to the bottom of the 900 m deep fjord, containing 200 m thick glaciomarine sediments. The relief together with the high precipitation, such as extreme rainfalls and heavy snowfalls, are responsible for slope processes of various kinds. In 2007 Bergen got 3024 mm of precipitation. Only three years ago Bergen experienced debris flows with fatal results (4 persons were killed in the autumn of 2005). Numerous slide scars and debris fans witness such active processes along the route. When we enter the heartland of the fjords we can demonstrate large old and new, as well as potential rock falls.

We will pass a few sites with Eemian sediments, witnessing relatively modest glacial erosion during the Weichselian glaciation. Fresh rock terraces produced by lacustrine cryoplanation will be demonstrated just 10 km north of the centre of Bergen. Here a 4 km^2 large area has a nearly horizontal bedrock surface that also demonstrates the main way the Norwegian strandflat was formed: Frost weathering close to an aquatic surface during periods when the coast was ice free.

The area of the first day’s excursion was mainly deglaciated during Preboreal time (11,500 – 10,200 years ago). The oblique glaci-isostatic rebound is demonstrated by the difference in the upper marine limit, being 45 m a.s.l. at the Bergen Airport and rising to 135 m a.s.l. at Flåm. The highlight of the day will be the two-hour boat trip on the Nærøyfjord where we will have our lunch as the scenic landscape of this UNESCO world heritage site passes leisurely by. From a new “skywalk” at 640 m a.s.l., we will study valley and fjord generations around the Aurlandfjord. Map: See Fig. 1.

Stop No 1: Bergen Airport, Flesland (60º 17.35’N, 05º 13.71’E)

“The Norwegian strandflat”

The strandflat near Bergen Airport is some 7 km wide, but not particularly level with heights from 40-60 m a.s.l. and small incised valleys. Some higher areas represent erosion remnants of older surfaces. (For description of the strandflat, see Regional geology).

Fjøsanger (no section, no stop, 60º 20.52’ N, 5º 19.83’ E)
On our way towards the centre of Bergen we pass the 84 m deep Nordåsvatnet fjord with anoxic bottom conditions (Paetzel & Schrader 1994, 1995). The fjord has a narrow and shallow inlet with a strong tidal current and earlier received sewage from the surrounding area. Closer to the town centre we pass the Fjøsanger locality where Mangerud et al. (1981) excavated a continuous Eemian - Early Weichselian sequence of shallow marine sediments and younger tills. The section is not longer accessible. The fact that these sediments are preserved in an area of relatively strong glacial erosion shows that glacial erosion is strongly selective.

Stop No 2: Bergen Bus Station (60º 23.33’ N, 5º 19.92’ E)
Stop outside the bus terminal (on the western side) to pick up participants that have arrived in Bergen a day ahead of the excursion.

Stop No 3: Nyborg, Åsane (60º 28.40’ N, 5º 20.22’ E)

Recently a “lacustrine strandflat” was discovered in the area around Nyborg, Åsane, just 10 km north of the centre of Bergen (Aarseth & Fossen 2004a). Here an area of 4 km^2 has a remarkably even bedrock surface surrounded by 3-500 m high hills (Fig. 9). The bedrock is generally covered by 1-3 m thick bogs. Measurements from the construction of a 4 km long canal (Fig. 10), as well as from road works and buildings in the area, together with measurements by probing and GPR (ground penetrating radar) unveiled an even bedrock surface 87 – 89 m a.s.l. This is 30-40 m above the general level of the marine strandflat in the vicinity. There are four small lakes, the largest, Langevatnet being 1.5 km long.

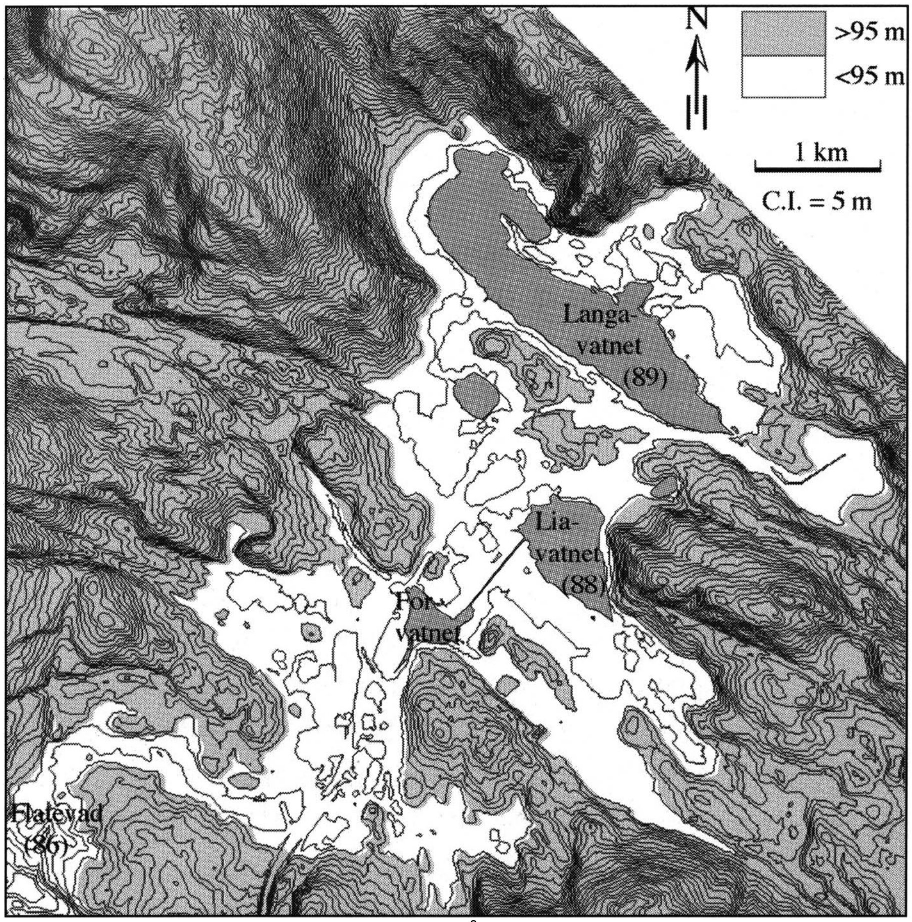

Fig. 9. Contour map of Midtbygda, Åsane showing areas below 95 m a.s.l., (white). (Aarseth & Fossen 2004a).

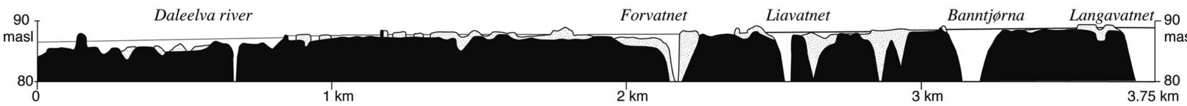

Fig. 10. Profile along the river through Midtbygda, Åsane showing the bedrock surface and thickness of organic sediment (Aarseth & Fossen 2004a).

Storavatnet, Osterøy (Not visited during the excursion)

On the neighbouring island Osterøy a bedrock terrace surrounds the 10 km^2 lake "Storavatnet" (Aarseth & Fossen 2004b). This lake was converted into a reservoir in 1920 and is therefore subject to frequent changes in water level. Along the shores a rock platform has developed at a mean altitude of 151.5 m a.s.l. The width of the platform is maximum 16 m. Frost weathering along the lake shores is suggested as the main process for the formation of the rock platform, and scouring by drifting ice is considered responsible for the removal of weathered material. On Osterøy the age of the rock platform is suggested to be maximum 5-6000 years based on the fact that the glacial rebound has not been oblique during this time.

The relatively rapid frost weathering along the shores of lakes, both on Osterøy and in Åsane is also partly due to climatic conditions. The Bergen area is located close to the Polar Front were the weather changes rapidly. Both Lake Storavatnet and the Nyborg area are located in topographic depressions were cold air collects during anticyclones in wintertime. Measurements of temperature inversions show differences of up to 10º from the basins to areas just 50 m higher. This normally occurs several times during the winter. During periods in between, cyclones bring mild temperatures, and lots of rain forces the ice covered lake level to rise.

The wide horizontal bedrock terrace in Åsane is thought to be older than Weichsel maximum. During interstadials this area was situated beyond the continental ice cap and frost weathering along the lake shore prevailed in wintertime, and most likely the bedrock surface was repeatedly being formed during these intervals. In some areas the bedrock is also extensively sheeted with sub-horizontal joints facilitating this kind of weathering.

Structurally controlled fjords

The bedrock structures and petrography has had an important role for the architecture of the fjord landscape in the Bergen area (Fig. 2 & 4). The topography mirrors the Caledonian structures of "The Bergen Arcs" with valleys and fjords being parallel to, or at right angles to the bedrock strike (Kolderup & Kolderup 1940). Some fjords have a direction closer to N-S. These fjords have steeper mountain sides controlled by nearly vertical coast-parallel fractures thought to have been formed during the opening of the North Sea basin in the Permian, Triassic and Jurassic. The best example of a fracture-fjord is the 30 km long and max. 400 m deep Veafjorden to the east of Osterøy island. This fjord is situated in Proterozoic gneiss just outside the rocks of the Bergen Arc System. On both sides of the fjord there is a relatively even mountain plateau at 600-800 m a.s.l.

The Bolstadfjord

From the N-S trending fjord Veafjorden, the E-16 turns towards the NE, following the main direction of the Caledonian orogenic belt. The Bolstadfjord is a "land-locked" fjord with anoxic bottom sediments (Strøm 1936; Taylor & Price 1983). This up to 160 m deep fjord is connected to the Osterfjord through a 1.5m deep threshold with a very strong tidal current. The fjord is also "base of erosion" for the Voss river system, the largest river system in the county of Hordaland, with a natural drainage area of 1500 km^2. The fresh- and brackish surface waters are trapping the heavier saltwater (20 ‰ salt) and thus preventing vertical circulation. Organic sediments are therefore being deposited on the fjord bottom and H_2S is found in the fjord at depths greater than 40-50 m.

The lower Voss valley

The 4 km long river below Lake Evanger (10 m a.s.l.) is named Bolstadelvi river and enters the sea at Bolstad. The upper marine terrace at Bolstadøyri is 63 m a.s.l Above Lake Evanger the river is called Vosso. The lower Vosso river flows on a narrow floodplain sometimes bordered by bedrock. Parts of the riverbed contain large boulders witnessing strong river currents during floods. Only one of the tributary rivers has been exploited for hydroelectricity, the rest is now protected.

Lake Vangsvatnet (47 m a.s.l.)

Both to the north and south of Voss the fjords penetrate further inland than Voss. The climate at Voss is even so more continental than along the fjords in Hardanger to the south and Sogn to the north. This is a result of the heat reservoir in the fjords. The natural outlet from Lake Vangsvatnet is rather narrow, and Voss town centre used to suffer frequent flooding which now is eliminated thanks to a lowered threshold. The lake consists of two basins, the easternmost (60 m deep) reaches below sea level

Development of the Voss river system in relation to the neighbouring fjords

The landscape at Voss is much less influenced by glacial erosion than the areas along the inner parts of the fjords in Hardanger and Sogn. The reason for this was suggested by Holtedahl (1975) to be due to the Vosso river being the original river that also drained the inner parts of Hardanger and Sogn. The preglacial rivers in the Sogn valley and the Hardanger valley drained larger areas than the Voss river system, Fig. 11 (Aarseth 2004).

Their water divides reached the main divide towards East Norway while the drainage area of the Vosso River was much smaller. During the early glaciations tributary glaciers coalesced and developed ice streams that drained through the larger river valleys. The confluence of tributary glaciers increased erosion and led to the formation of deep fjords. Thus the base of erosion was brought far inland in Sogn and Hardanger. The interglacial and interstadial rivers in combination with glacial erosion could capture some of the upper tributaries of the Voss river system. This will be studied at the “classic” locality at Stalheim.

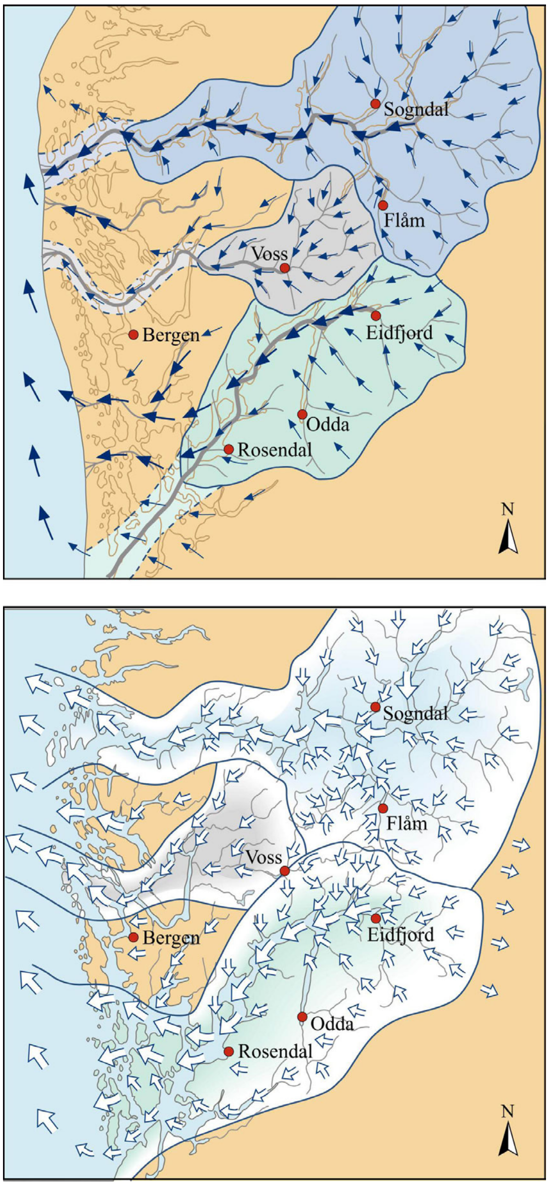

Fig. 11. Reconstruction of the preglacial drainage areas and rivers of the three districts: Hardanger, Voss and Sogn. Blue arrows indicate how early glaciers followed the valleys (upper). The lower map shows ice flow and local glacial divides during the Weichselian (last) glaciation (Aarseth 2004).

Stop No 4: Bordalsgjelet, Voss (60º 36.75’ N, 6º 25.37’ E)

The tributary valleys leading down to Lake Vangsvatnet are hanging valleys. The most spectacular is the Bordalen valley with prominent ledges illustrating the former valley floor. Bordalsgjelet canyon is a “valley of adjustment” connecting the former valley floor in Bordalen valley to Lake Vangsvatnet with a difference in elevation of 250 m. Parts of this 5 km long canyon is accessible to the public along a narrow nature trail into the canyon. The canyon is developed in phyllites and contains many large and small, half and whole potholes. The bridge crossing the canyon is 32 m long and the canyon under the bridge 32 m deep. A poster describing the geology, ornithology and road history is placed at the parking lot as well as under the bridge.

At the lower end of the canyon two glacial accumulations are found. To the west a fan composed of glaciofluvial material reaches 160 m a.s.l. This is thought to represent a subglacial accumulation, and lumps of very hard till were found in a gravel pit (Skreden 1967). The canyon may have been filled with till during parts of the Weichselian glaciation.

The other deposit is a marine delta with the highest level marking the upper marine limit in the area, 97 m a.s.l. when a fjord reached all the way to Voss.

Vossestrand

The tributary river from the north is named Strondaelvi and comes from Lake Oppheimsvatnet. The river flows parallel to the E-16. Just north of Voss thick accumulations of till was plastered on the western side of the valley (Mangerud & Skreden 1972). A section containing Eemian sediments was found along the road in the Vinjadalen valley, just below the outlet of Lake Oppheimsvatnet (Eide & Sindre 1985). Here a 2 cm thick layer of very compact peat was found containing pollen from *picea* (spruce). The spruce trees growing in the area today are planted. The organic layer lay beneath a thick sequence of laminated glaciolacustrine sediments deposited between the glacier coming from Myrkdalen and the watershed to the east. The direction of this part of the valley (NNW-SSE) is normal to the direction of the main glacial movement. The fine sediments capping the peat deposit have also protected it from erosion. A thick till is presently covering the sediment sequence.

Stop No 5: Stalheim Hotel terrace, Lake Oppheimsvatnet and Stalheim

(Stalheim Hotel terrace: 60º 50.01’ N, 6º 40.87’ E)

The 5 km long Lake Oppheimsvatnet (332 m a.s.l.) is situated close to the local water divide just east of the lake (Haugsvik at 333 m a.s.l.). Looking towards east from the western shores of this lake we can see several valleys pointing towards the lake, but their rivers are changing direction and are now captured by the valley leading down to the Nærøyfjord at Gudvangen. This is even better studied at the classical site Stalheim (Reusch 1901; Ahlman 1919). Here Reusch described “hook-valleys” (the norwegian word “agnordal” which means fish-hook valley: a valley with an acute angle). The view from the terrace at Stalheim tourist hotel gives an instructive demonstration of river piracy. Here the backward erosion, mainly of glacial origin, is responsible for capturing rivers from several tributary valleys: The Jordalen and Brekkedalen valley on the left (north) and Øvsthusdalen and Brandsetdalen valleys in the background on the right (south). In addition to changes in river courses (Fig.12) the valley shoulders of the Nærøydalen valley itself clearly demonstrates the different valley generations. The Nærøydalen valley ends at the waterfall Stalheimfossen (126 m high). This valley head is a typical glacial landform found in many western Norwegian valleys.

Nærøydalen valley

The Nærøydalen valley and the Nærøyfjord together with the Geirangerfjord area were listed as UNESCO world heritage sites in 2005. This was mainly due to the geology and spectacular geomorphology of the areas. The high relief and the narrow valley and fjord make a strong impression on visitors. Maximum visual relief is seen at the quay at Gudvangen where the edge of the cliff is 1420 m a.s.l. The fjord and valley is cut far into the paleic surface resulting in extreme hanging valleys with waterfalls seemingly coming from the sky (Fig. 13). The active exogenic processes are clearly demonstrated both from the steep mountain slopes and along the Nærøydalen river. There are plans to increase the ongoing mining operation for anorthosite, but no final decision has yet been made. The plans involve shipment during the night from an artificial cavern inside the mountain near Gudvangen. The anorthosite is a future raw material for aluminium production, but is now used in a variety of ways due to its whiteness.

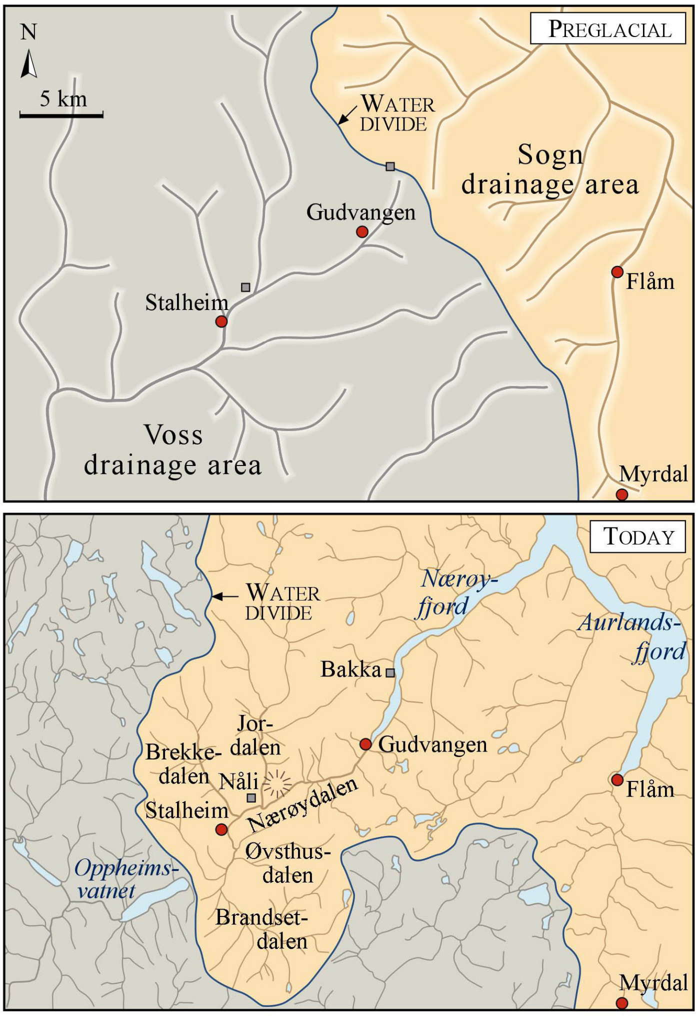

Fig. 12. Reconstruction of the preglacial drainage pattern in the Voss – Sogn area and the change in the water divide. (From Norwegian Nomination 2004: The West Norwegian Fjords).

Fig. 13. The waterfall "Kjelsfossen" just east of Gudvangen village. Photo I. Aarseth 2004.

Stop No 6: Gudvangen (60º 52.77' N, 6º 40.87' E)

The small farms in Gudvangen have long suffered from strong winds generated by snow avalanches. Stone fences were built long ago on the lower sides of the houses to protect them from avalanche winds from the opposite side of the valley! Before construction of the new protection ramparts in 1998 measurements were carried out on the snow avalanches as well as on the stability of the avalanche fans. Two large artificial snow avalanche ramparts

Langageiti and *Nautagrovi* were constructed to protect the village and the harbour at Gudvangen. Langageiti is 570 m long and max 13 m high. Here the slide volume may reach 50.000 m^3 with a velocity of up to 40-50 m/s in the steepest part of the slide path. The wind pressure may reach 1 Ton/m^2 and the pressure from the snow itself 12 Tons/m^2. At the northernmost avalanche rampart, Nautagrovi (310 m long, Fig. 14) the slide volume may reach 100.000 m^3 and the slide velocity 30 m/s. Wind pressure is measured to 0.6 Ton/m^2 and the pressure from the loose snow itself as much as 6.8 Tons/m^2 at the Hotel (Project brochure).

Fig.14. Snow avalanche rampart "Nautagrovi" just above the hotel and the quay at Gudvangen. Photo: I. Aarseth March 2007.

Cruise on the Nærøyfjord and the Aurlandfjord

On this 30 km long cruise we will be able to see geomorphologic landforms and processes as well as historical land use along the fjord. Very little detailed geomorphologic studies have been carried out in this area. Sindre (1973) suggested that the preglacial water divide was situated on the mountains above Bakka. The mountain peaks in this area were higher than to the southwest and northeast demonstrating a possible former dome. The inner fjord basin is relatively shallow with only 70 m water depth inside the 11 m deep threshold complex at Bakka where several shoals make a tricky sailing path for the cruise ships. The threshold is made up of terminal moraines as well as rock falls (Sindre 1973). Further out the fjord becomes gradually deeper with a 150 m step down to the Aurlandfjord which is 500 m deep at the intersection. Along the Aurlandfjord valley ledges are found representing older valley generations. The old farms at Stigen (348 m a.s.l.) and Nedbergo (545 m a.s.l.) are located on two such sloping ledges. Undredal valley is the largest tributary valley. Here sloping terraces indicate a former valley fill of glacifluvial material up to more than 150 m a.s.l. The village is well known for its production of goat cheese. Goats are the animals best adapted to this kind of topography!

The sediment basin just outside Cape Flenes is 419 m deep and has a sediment thickness of 200 m. The highest mountain to the NE, Blåskavlen is 1809 m high. This gives a bedrock relief of 2400 m in this area. The 1000 m high mountain slopes to the north are striking contrasts to the even plateau just above the escarpment. The slopes contain several ravines and gorges pointing to a strong fluvial activity during interstadials as well as interglacials. At Cape Flenes the fjord turns abruptly toward SSW, eroded along the thrust plane between the Jotun Nappe to the west and the underlying Paleozoic phyllites to the east.

The Aurlandsdalen valley is eroded into the Precambrian basement. At Cape Otnes the contact between the migmatitic gneiss and the phyllites clearly demonstrates a shear zone with strongly foliated gneissic rocks below the contact. Here loose blocks of gneiss rest on a steep surface (Hardangerfjord Shear Zone) just above the newly built houses in the area.

The asymmetry of this part of the fjord as well as the lower part of the Flåm valley is due the westward dipping bedrock units. Large rockslides are localised 3 km north of the head of the fjord as well as on the eastern mountainside, Fig. 15 (Braathen et al. 2004). Open joints as well as talus blocks are now monitored on the phyllite slopes above the main road (E-16). A total volume of 900-1500 million m^3 phyllite is considered unstable in the Aurland – Flåm area. Talus creep has been observed at several locations and some minor talus avalanches have taken place in recent years in connection with severe rainfall. A basal date of a sediment core from the inner part of the Aurlandfjord indicates an age of approximately 3000 cal. years for the large rock avalanche in the fjord (Bøe et al. 2004).

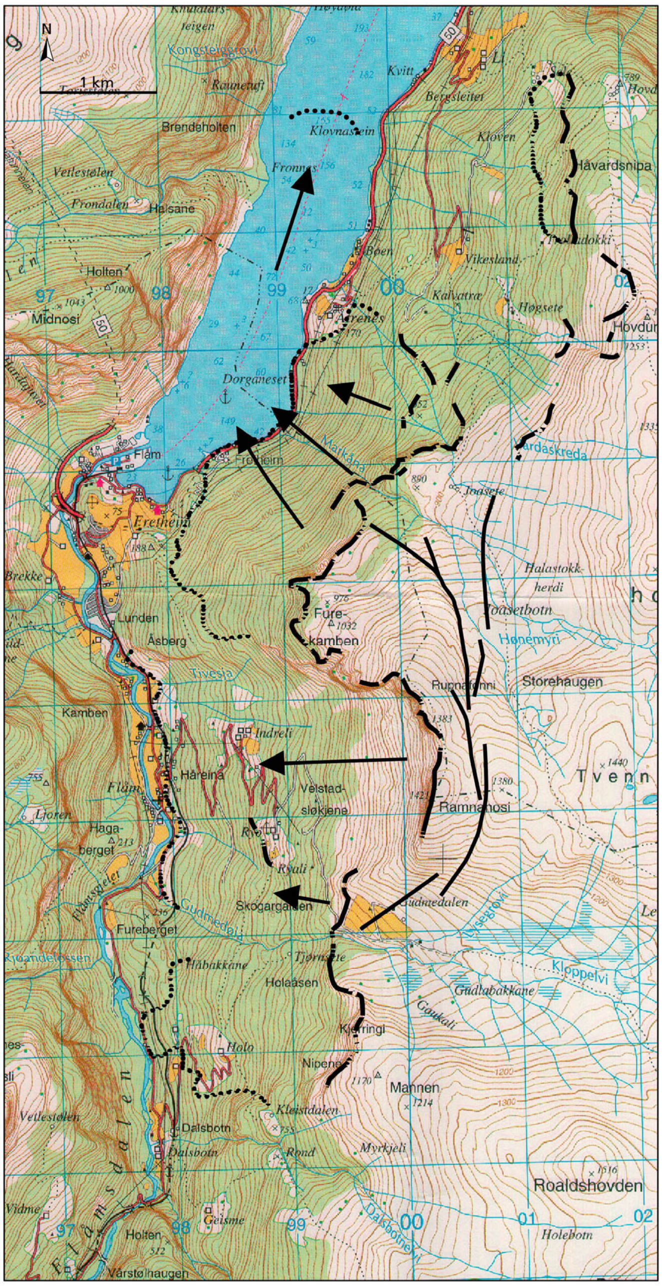

Fig. 15. The inner part of the Aurlandsfjord and the lower Flåm valley. Old rock slide tongues and open joints representing large potential rock slides (Braathen et al. 2004).

Stop No 7: Flåm church (60º 50.22’ N, 7º 07.30’ E)

At a stop near Flåm church we will demonstrate large Holocene slide tongues and glacial frontal deposits. Horizontal terraces on both sides of the river indicate an upper marine limit of 135 m a.s.l., the highest marine limit in western Norway, (Fig. 16). In the distance we can see the contact between the Jotun Nappe and the underlying phyllites.

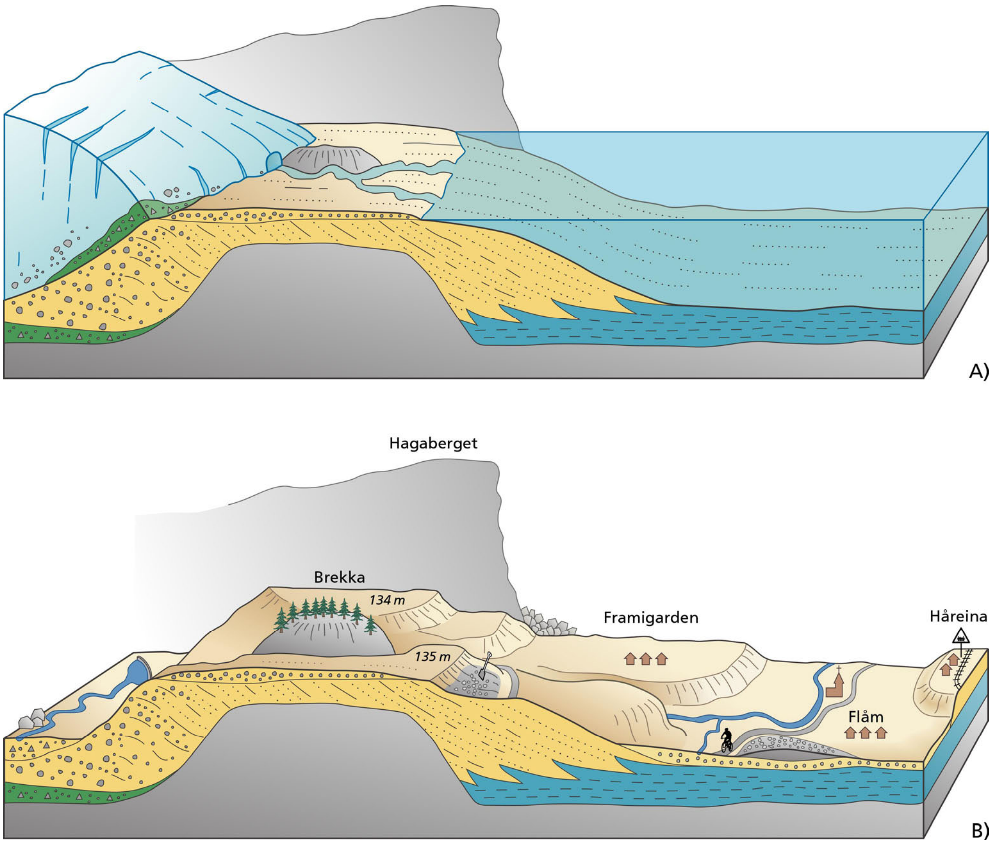

Fig. 16. Upper: Model of the deposition of a typical Preboreal icemarginal delta at the inner fjords of western Norway; example from the Flåm valley. Lower: Present day terraces with the upper marine limit at Flåm (135 m a.s.l), Aarseth et al.(2006).

Aurlandsvangen village.

The glacial geology of the Aurlandsdalen valley shows a series of recessional terminal as well as lateral moraines deposited during deglaciation in the Early Preboreal (Bergstrøm 1975). There will be no time to study these apart from the view from the boat and the bus window.

Stop No 8: Stegastein “sky walk”, 640 m a.s.l. (60º 54.50’ N, 7º 12.75’ E)

A new “skywalk” was opened in 2006 as part of the “National Tourist Route” Aurlandsvegen road. This rest area is situated at the transition between the Proterozoic basement (gneissic rocks) and the overlying unit of Older Paleozoic phyllite. North of here and across the fjord we can see rocks of the Caledonian Jotun Nappe (hyperstenmonzonitt – mangeritt).

The skywalk gives a splendid view of the mountains and fjord (Fig. 17). The view can be used to demonstrate the geomorphologic development of several valley generations as well as younger tributary gorges adapting to the present sea level. The upper gentle surface represents the paleic (old) surface formed after the epeirogenic uplift of the Scandinavian peninsula during Cenozoic time. An early uplift took place in connection with the opening of the Norwegian Sea between Norway and Greenland that started 56 mill years ago. A later uplift took place possibly in Miocene – Pliocene. Subsequently series of younger fluvial valley generations developed before the Quaternary glaciations intensified the erosion that led to the formation of fjords. From here we see excellent examples of “truncated spurs” where glacial erosion has straightened former winding fluvial valleys.

Fig. 17. View from the “skywalk” at “Stegastein”, Aurland (640 m a.s.l.; Photo: I. Aarseth 2007).

Stop No 9: Nalfarhøgdi 1300 m a.s.l. (60º 55.50’ N, 7º 18.59’ E)

The so called “snow road” leads further up a hanging valley (Kvammadal), passing mountain farms situated on thick fertile till deposits, and reaches the mountain plateau Nalfarhøgdi at 1300 m a.s.l. (**Photo stop**). This is a northern extension of the Hardangervidda mountain plateau, the largest in northern Europe. Several lakes are strewn across the plateau and mountaintops represent remnants of the over trusted Jotun Nappe, the most prominent being Hornsnipa, a 1692 m high peak just west of the road. The plateau is the main drainage area for large hydroelectric plants with reservoirs at 1400-1450 m a.s.l. The environment along the next 10 km of the road on the plateau changes from a stony and barren landscape on Precambrian migmatitic gneiss to green pastures on phyllites. Periglacial frost phenomena are common along the road. The road leads down to Lærdalsfjord, a branch of the Sognefjord, through the V-shaped Erdalen valley. Below 130 m a.s.l. the valley has sloping sediment terraces deposited during the deglaciation.

The Sognefjord.

This excursion only touches the innermost part of the nearly 200 km long and up to 1308 m deep Sognefjord + 200 m of deglaciation sediments (Aarseth 1997). Nesje & Whillans (1994) discussed the formation of the fjord and emphasised fluvial down-slope action during interstadials and interglacials being responsible for the formation of numerous V-shaped valleys and slide scars along the fjord. Glacial processes, according to Nesje & Whillans

(1994), were only responsible for removal of the erosion products and deepening of the fjord below sea level. They proposed a model for the formation of the fjord (Fig. 18).

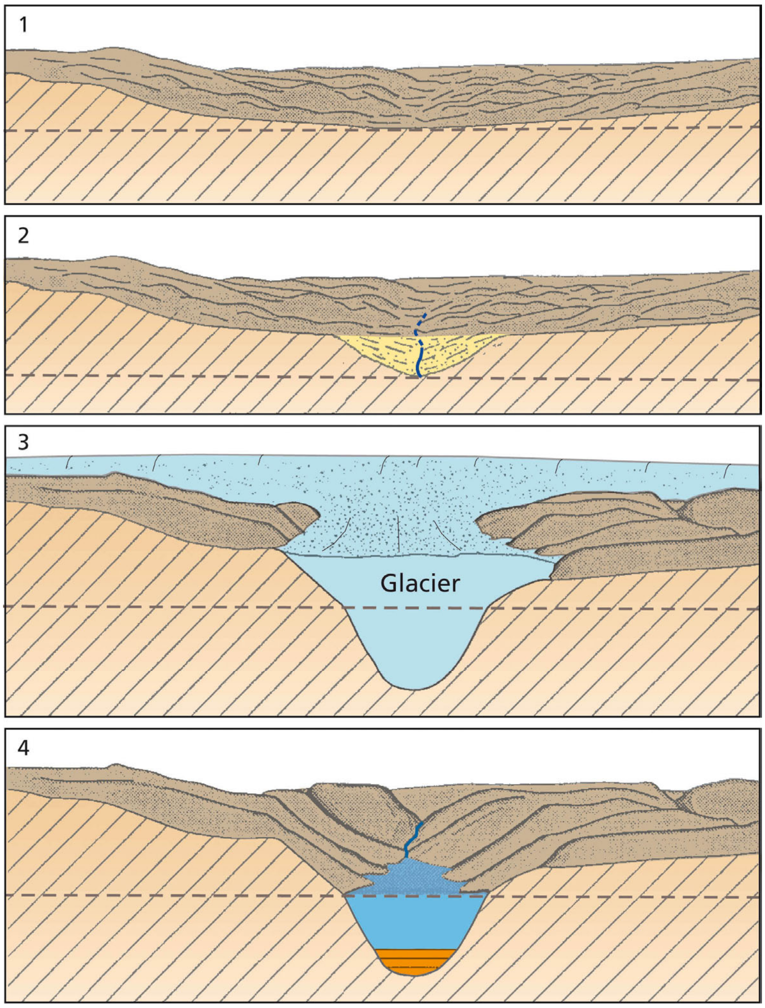

Fig. 18. Phases in proposed formation history of the Sognefjord. Sea level is represented by the dashed line. 1. Paleic landscape, 2. Fluvial stage, 3. Glacial stage. 4. Present stage. (Slightly modified from Nesje & Whillans 1994).

Stop No 10: Fodnes: Ferry Fodnes – Mannhiller

(Fodnes: 61º 08.95’ N, 7º 23.18’ E).

At the ferry crossing Fodnes – Mannhiller, the Sognefjord is 824 m deep. 10 km west of here a record bedrock relief is found. The vertical distance from the edge of the mountain Bleia at 1660 m to the 920 m deep fjord with 200 m thick sediments is approximately 2800 m over a horizontal distance of 4000 m. This is almost twice the relief of the famous Grand Canyon in the US. A little further west the former bottom of the fluvial valley can be found as broad valley ledges on both sides of the fjord, and on the north side the local airport is located at 500 m a.s.l. (Fig.19, Aarseth 1980).

The road towards Sogndal leads through a paleo-valley (at Kaupangerskogen industrial area). Just before Sogndal village a bridge crosses a tidal current with an 8 m deep threshold to the Barsnes fjord. This fjord has anoxic fjord sediments in the 80 m and 66 m deep fjord basins (Paetzel & Schrader 1992).

<u>Evening meal and stay overnight at Hofslund hotel, Sogndal, phone: (+ 47) 57 62 76 00</u>

Hofslund hotel: (61° 13.91' N, 7° 06.49' E)

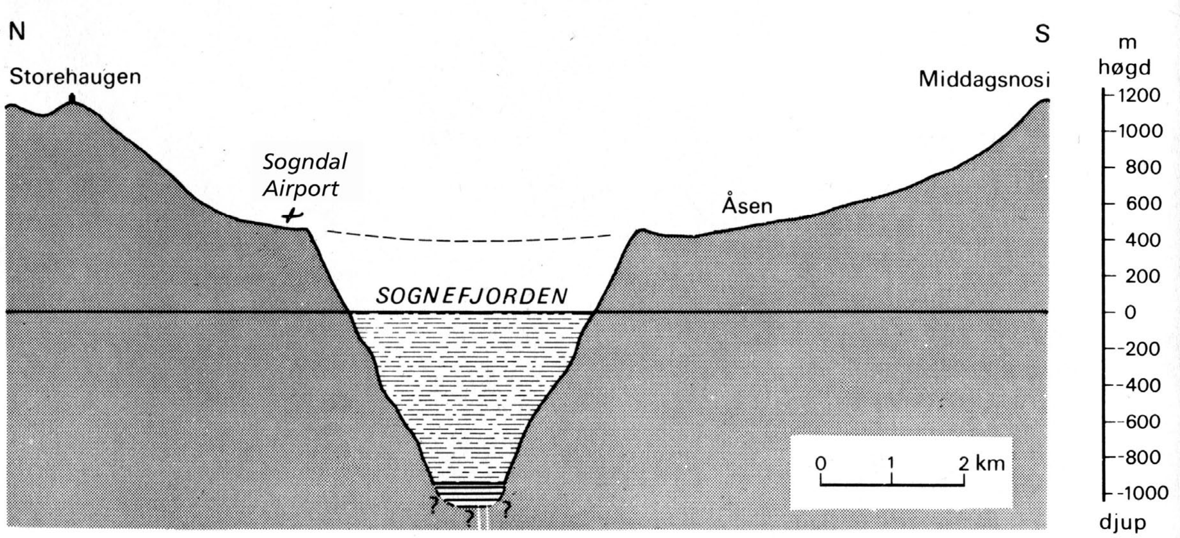

Fig. 19. Profile across the Sognefjord 15 km W of the ferry crossing Fodnes – Mannhiller. Sogndal airport is situated on the preglacial valley shoulder (Aarseth 1980.)

Day 2: Friday 15th August: Sogndal – Norfjordeid

Introduction

Today we will drive along the "Geo nature route" (From Sogndal to the E39 at Skei) in front of, as well as under, the Jostedal Glacier. Here geology has been made accessible for the public in the form of posters and models at several scenic rest areas.
The Norwegian Glacier Museum in Fjærland will be visited. The museum was opened in 1991. A new climate exhibition was opened the summer of 2007. In Fjærland we will also visit two glaciers: Supphellebreen and Bøyabreen (lunch at the café). Fjærland is surrounded by glaciers with a great variety in size from small cirque glaciers to Jostedalsbreen, the largest ice cap on mainland Europe. Jostedalsbreen is approximately 80 km long and covers an area of 487 km^2. The highest elevation of Jostedalsbreen is 1957 m (Høgste Breakulen), and the lowest altitude is the tongue of Supphellebreen at 60 m. More than 30 named outlet glaciers flow from the ice cap.

On the north side of the 6,4 km long Fjærland tunnel, the "Georoad" takes us along the Kjøsnesfjord (a tributary lake to Lake Jølstervatn) with well developed sheeting on the opposite mountain side. The outlet of the 22 km long Lake Jølstervatn has shifted due to oblique isostatic uplift during Holocene. Now it drains towards the SW.

As we approach the Nordfjord area we will get closer to the terminal moraines of the Younger Dryas re-advance. During this phase the main glacier was split in three lobes, the southernmost in the Gloppen fjord, the middle in the Nordfjord itself and the northernmost in the Lake Hornindalsvatn east of Nordfjordeid. Large moraines are visible both below and above the upper marine limit, and seismic surveys in the fjords have revealed the gigantic moraines as well as up to 350 m thick sequences of distal glacimarine sediments.

Sogndal valley

The river in the Sogndal valley drains an area with up to 1600 m high mountains with some smaller cirque glaciers. The lower parts of the valley are relatively steep and lack a flood plain above the delta area. The upper parts have a few lakes of which Lake Dalavatnet (398 m a.s.l.) is the largest. During road work a section containing Middle Weichselian interstadial sediments was found in the valley (Aa & Sønstegaard 2001). At 380 m a.s.l. a 20 cm thick bed with gyttja silt is overlain by 3 m of glaciolacustrine silt and 5-6 m of till. Pollen from the gyttja silt indicates an open treeless landscape. Two radiocarbon age estimates indicate an interstadial of Middle Weichselian age.

Stop No 1, Vatnasete rest area (61º 19.97’ N, 6º 55.51’ E)

Along this “Geo Nature Route” several new rest areas are supplied with information on the geology and geomorphology of the area. There will be short stops on both sides of the 6746 m long Frudalstunnel: Vatnasete and Berge. At Vatasete history since the ice age is illustrated as a walk back through history with historic and prehistoric landmarks engraved on the “stepping stones”. In addition the formation of lacustrine deltas are shown. Snow avalanche protection ramparts are also built along the road in this area (Nesje et al. 1994).

Stop No 2: Berge rest area (61º 22.86’ N, 6º 45.47’ E)

The rest area at the western end of the tunnel overlooks the Fjærlandsfjord and explains the formation of fjords and the deposition of fjord sediments. The sedimentation rates in this fjord are relatively high. ^{210}Pb measurements gave rates of 4 mm year^{-1} just 2.2 km from the delta front over the last 90 years. 6.5 km from the delta the rate decreases to 1.2 mm year^{-1} (Vangsnes 1981; Aarseth 1988). The delta front had prograded at an average rate of 2.5 m year^{-1} in the period 1830-1950 (Mundal 1953).

Stop No 3: The Norwegian Glacier Museum, Fjærland (61º 25.39’ N, 6º 45.67’ E).

The Norwegian Glacier Museum is a non-profit making foundation established by the International Glaciological Society, Norwegian Mountain Touring Association, Norwegian Water Resources and Energy Directorate, Norwegian Polar Institute, Sogn og Fjordane Regional College, the University of Bergen and the University of Oslo. The museum was opened by Her Majesty Queen Sonja in the summer of 1991 and attracts 50-60.000 visitors every year. The exhibitions deal with 24 themes in four main categories. They show among other things how glaciers build up, how they shape the landscape, and why they play an important role in the search for knowledge about the past and future climate.

The “Ulltveit-Moe Climate Centre” was opened in the summer of 2007 by former US vice president Walter Mondale. The centre is part of the museum and takes you on a journey through time, from the creation of the earth, through the last ice age and finally to the year 2100. The first part of the exhibition introduces natural climate changes. The second part of the exhibition deals with our common future.

Stop No 4: Supphellebreen glacier (61º 27.75’ N, 6º 49.28’ E)

Supphellebreen is a 0.1 km^2 regenerated glacier extending from 320 to 60 m a.s.l., making this the lowest-lying glacier in southern Norway, (Fig. 20). The glacier is nourished by ice breaking off from the ~50 m high front of Flatbreen, which covers 11.8 km^2 and extends from

1740-720 m elevation. Supphellebreen reached its maximum post-glacial extent around AD 1750, about 800 m further down the valley from the present front. Even at that time Supphellebreen was a regenerated glacier. Investigations of Supphellebreen from 1963 to1967 showed that about 2 million tons of ice broke off from Flatbreen annually. This was equal to adding a 15 m thick ice layer to Supphellebreen, or expressed in other terms: if the ice was drink-size ice cubes and placed end to end, they would stretch 50 times around the Earth (Orheim 1970). The flow rate at the terminus of Flatbreen was approximately 2 m/day, and the response time at the snout of Flatbreen to changes in its mass balance was 2-3 years in that period.

Fig. 20. Supphellebreen regenerated glacier, Fjærland. (Photo: I. Aarseth 2007).

Stop No 5: Øygard debris fan (61º 27.25’ N, 6º 48.10’ E).

Every summer a lake is dammed between the terminal moraine complex of Flatbreen and the glacier terminus. The lake normally empties through channels under the glacier over a short period of time (hours). During the early summer of 2004 the lake grew to reach the top of the moraine resulting in a Jökulhlaup event on May 8th (Fig. 21). The melt water flushed ~ 1000 vertical meters down the tributary valley as a major debris flow. On the floodplain it destroyed a vast cultivated area. The road to the glacier crosses the new debris fan with up to car-sized boulders (Breien et al. in press).

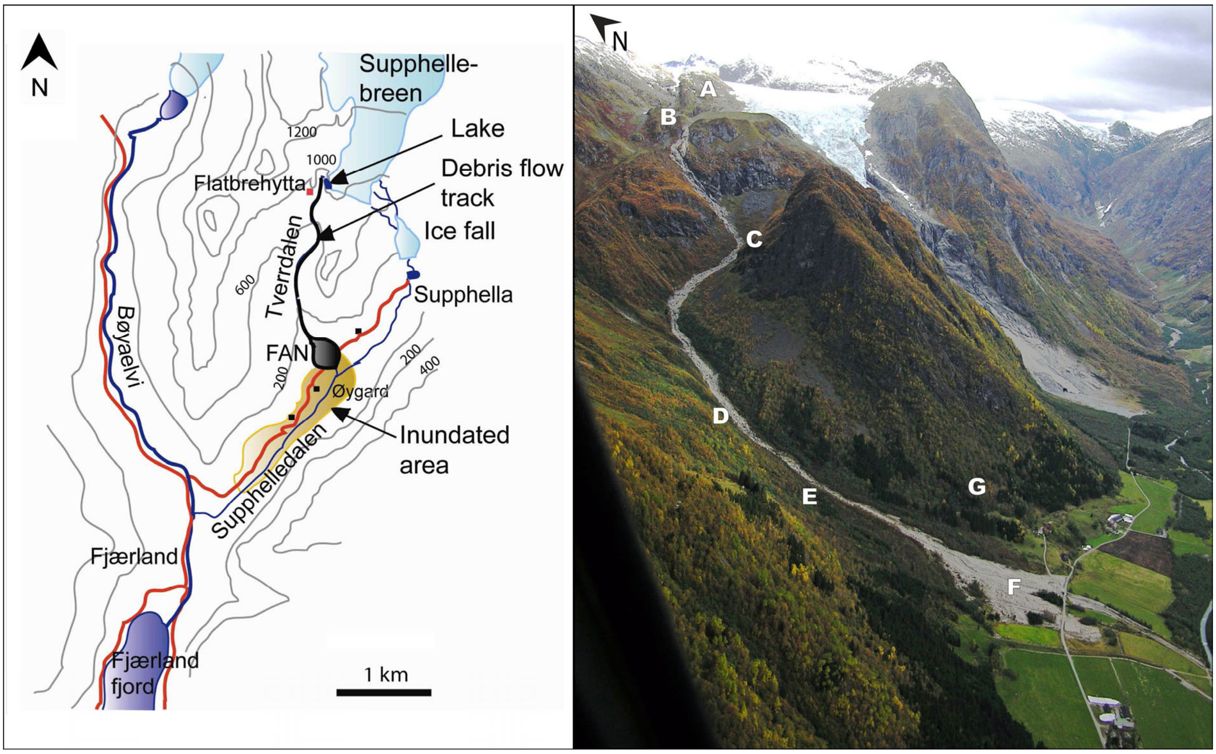

Fig. 21. Map of the Øygard jökulhlaup 2004, and photo of the flow path. (Photo: A. Elverhøi 2007, from Breien et al., in press).

Stop No 6: Bøyabreen glacier (61º 29.08' N, 6º 45.28' E)

Bøyabreen is 5.7 km long and covers an area of 13.9 km^2. The skyline of the glacier is at 1000 m elevation. The glacier extends down to about 490 m a.s.l. Below the terminus there is a small regenerated glacier, Fig. 22. Due to a significant glacial advance in the late 1990s, the front of Bøyabreen was connected to the regenerated glacier on the right-hand side. In Bøyadalen, several marginal moraines document minor advances/halts during glacier recession after the 'Little Ice Age' maximum in the middle of the 18th century. The largest terminal moraine M1, 2 km below the glacier margin, dates to the Early Holocene (Fig. 23). The younger moraines M2 and M3 were estimated to be of Late Holocene age by the use of Schmidt hammer measurements (Aa & Sjåstad 2000). Posters in the area show the Holocene history of the glacier. Before looking more closely at Bøyabreen, lunch will be served at the "Glacier Lake Cabin" in front of the glacier.

Fig. 22. Bøyabreen glacier, 1991, 1995 and 2007.(Photos: I. Aarseth).

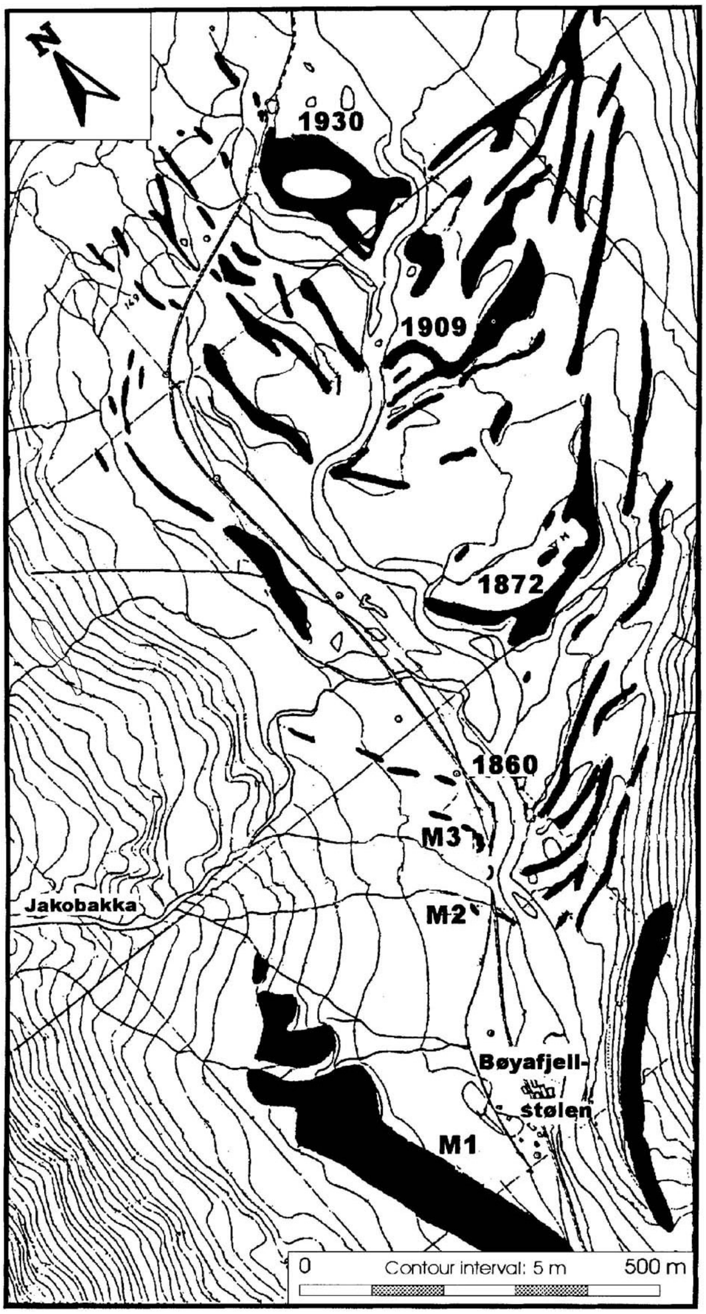

Fig. 23. Map of the moraines in front of the Bøyabreen glacier. (Aa & Sjåstad 2000).

Stop No 7: Lake Kjøsnesfjord (61º 31.98' N, 6º 33.94' E)

On the opposite side of the Fjærland tunnel the road pass along Lake Kjøsnesfjord. Due to frequent snow avalanches, new concrete and rock tunnels have been built to protect the most exposed areas along the road. On the opposite side of the lake the gneissic rocks have developed strong sheeting as joints subparallel to the steep rock surfaces. In Norway this is called "valley joints". Sheeting occurs when the gneissic rocks are less foliated than normal. Snow avalanches transport the weathered debris to the foothills.

Stop No 8: Lake Jølstravatnet, Skei, 207 m a.s.l. (61º 34.35’ N, 6º 28.80’ E)

This 22 km long and 40 km^2 large lake has a maximum depth of 233 m. Several tributary valleys enter the lake as hanging valleys. The Holocene drainage history of the lake shows that the outlet has changed from the northeast to the southwest end (Klakegg & Rye 1990). The study is based on investigations of shore levels as well as lake sediments.
The change was caused by faster uplift at the eastern end. The lake history is subdivided into five phases. During phase I (-9500 BP) Lake Jølstravatnet was an ice-dammed lake. During phases II and III (9500 – 7500 BP) there was an eastern outlet, and glaci-isostatic tilting caused a transgression in the western part of the lake until the lake level rose to the threshold at the western end. Phase IV (7500 – 6000 BP) was characterised by outlets at both ends of the lake. The emergence of the eastern outlet (ca. 6000 BP) marks the transition to Phase V, i.e. the present situation (Fig. 24).

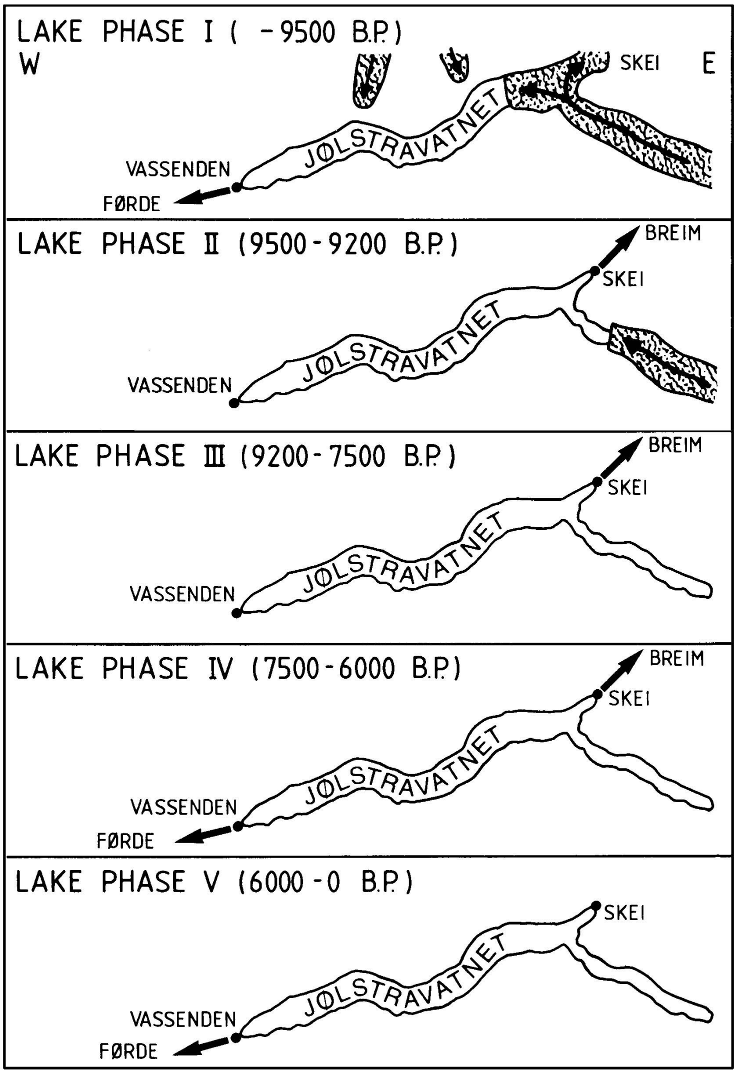

Fig. 24. Five phases of the outlet drainage of Lake Jølstravatn (Klakegg et al. 1990).

Våtedalen valley

Just north of the village of Skei the road leaves the former river valley from Lake Jølstervatn and leads through the narrow and picturesque N-S trending Våtedalen valley. The river flowing slowly through the valley drops only 10 m in several km, and during the summer it carries a load of rock flower from this part of Jostedalsbreen glacier. Large talus cones make a contrast to the meandering river on the flood plain and demonstrate that active frost weathering is taking place on the steep valley sides that are parallel to the N-S joint system.

Gloppen fjord

In the Nordfjord area the glacier split up into three glacier tongues during the Younger Dryas ice advance. In this area it is called the Nor stadial (Fig. 25; Fareth 1987). The southernmost branch advanced a short distance out into the Gloppen fjord outside Sandane village. Large lateral moraines were deposited on the SW side of the fjord at Rygg (which means “Ridge”). Seismic profiles reveals two ice push phases in this fjord, Fig. 26, Aarseth et al. (1997). On the Quaternary map three moraine ridges are distinguished (Klakegg & Nordahl-Olsen 1985, 1986).

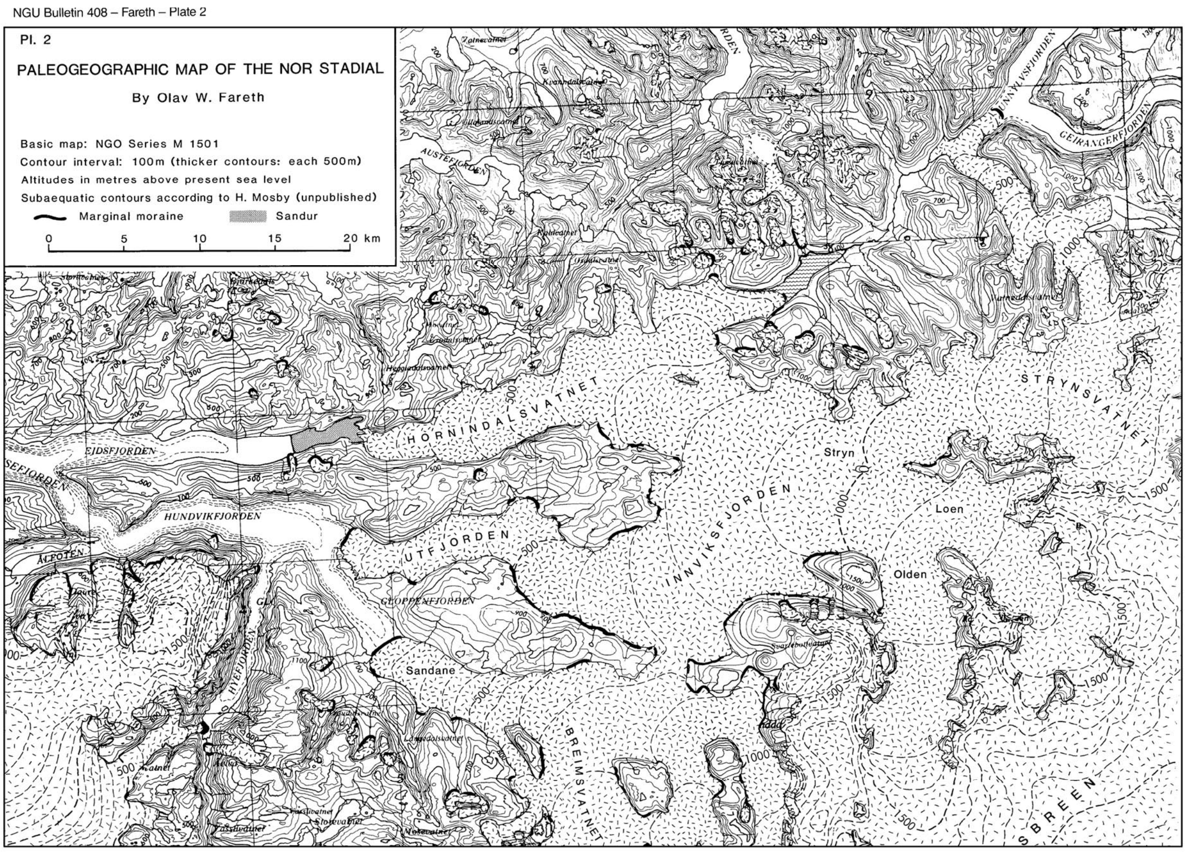

Fig. 25. Paleogeographic map of the Younger Dryas Nor stadial in the Nordfjord area. (Modified from Fareth 1987).

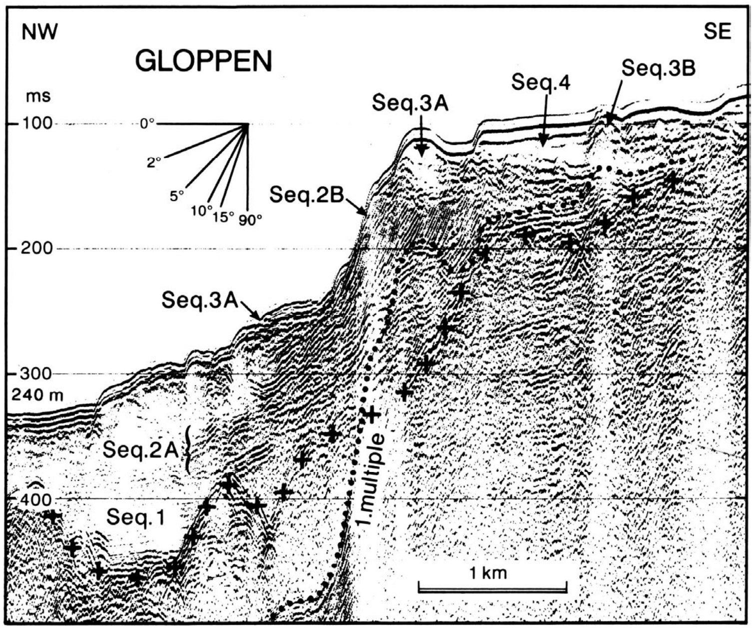

Fig. 26. Seismic profile (Sparker) of the terminal moraines and distal sediments deposited during the Younger Dryas in the Gloppen fjord. Two glacial readvances can be distinguished (Aarseth et al. 1997). +++ indicate acoustic basement (bedrock).

Stop No 9: Føleide, Gloppen (61° 49.57' N 06° 10.48' E)
The glacier tongue in the main fjord reached a low pass on the south side and a prominent lateral moraine was deposited at Føleide (Fareth 1987). It now constitutes the local watershed between the main fjord and the Gloppen fjord with a saddle point at 197 m a.s.l. Several farms are located on the moraine, but still we get the impression of a large ridge. The lateral drainage from the glacier has cut a valley through thick older till.

Stop No 10: Vereide, Gloppen (61° 48.62' N 06° 09.21' E)
The meltwater river reached the Gloppen fjord where it deposited a delta at Vereide, 69 m above the present sea level (Fareth 1987). This is the upper marine limit in this area.

Stop No 11: Anda Ferry Anda – Lote. Anda-Lote moraine threshold
(Anda: 61º 48.62’ N, 6º 09.21’ E)

During the Younger Dryas ice advance the fjord glacier terminated at the location of the present ferry crossing from Anda to Lote. The bottom topography reveals a 100 m high ridge that constitutes a fjord threshold with a saddle depth of 130 m. Seismic profiles (Sparker) penetrated a 200 m thick sediment sequence (Fig. 27; Aarseth et al. 1997). The moraine complex consists of a lower sequence of transparent glaciomarine sediments overlain by glacial ice push sediments. Above this is a 50-100 m thick sequence of glaciofluvial foreset beds, dipping ~15º to the west. The top of the ridge is composed of a 50 m thick compact till illustrating the last ice push. This ice oscillation could be due to reduced calving as the water depth decreased when the foreset beds were deposited (Aarseth et al. 1997).

On the way down from the Lote tunnel to Nordfjordeid village the road crosses a small terminal moraine deposited by a small cirque glacier on the north-sloping valley side.

Evening meal and stay overnight at Rica Partner Nordfjord hotel. (+ 47) 57 86 04 33
Nordfjord hotel: (61º 54.29’ N, 5º 59.77’ E)

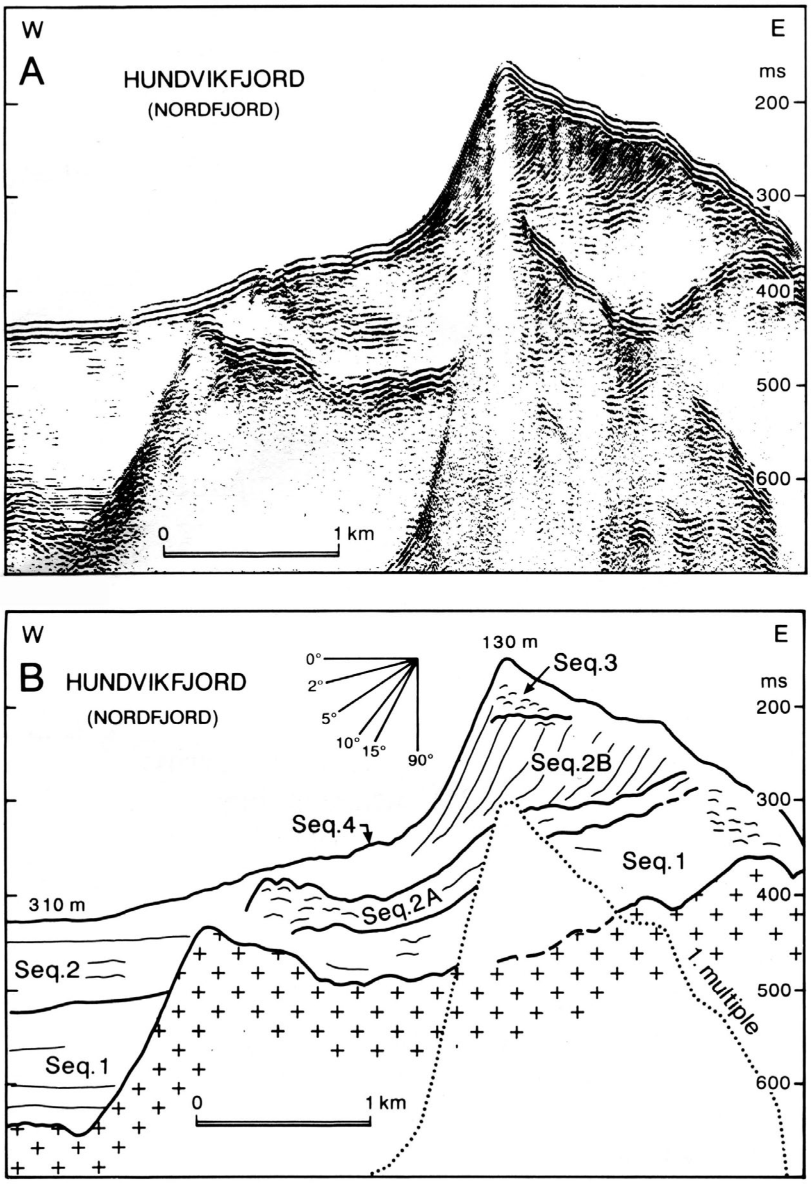

Fig. 27. Seismic profile (Sparker) across the Younger Dryas terminal moraine in the main Nordfjord. (Aarseth et al. 1997).

Day 3: Saturday 16th August: Nordfjordeid – Vågsøy –Nordfjordeid

Introduction

On the third day of the excursion we will focus on glacial/non-glacial geomorphology, land-ocean interactions, regeneration of Younger Dryas glaciers and coastal processes. There will be a quite long bus trip along the northern shore of Nordfjord to localities on the coastal island of Vågsøy. During the bus trip, there will be discussions on long-term development of the west-Norwegian landscape and fjords.

As discussed in the introduction to the excursion, summit areas in western Norway generally are unmodified by glacial erosion and probably are of significant (Neogene) age. However, also minor “pockets” of sediments and non-glacial landforms have been described on the Nordfjord coast (Roaldset et al., 1982; Longva et al., 1983). An overview of “relict” non-glacial landforms and saprolites in relation to a reconstructed LGM ice profile can be seen in Fig. 28. These localities have attracted the attention of researchers because they indicate very limited (local) glacial erosion and may thus give insights into past glacial- and non-glacial processes.

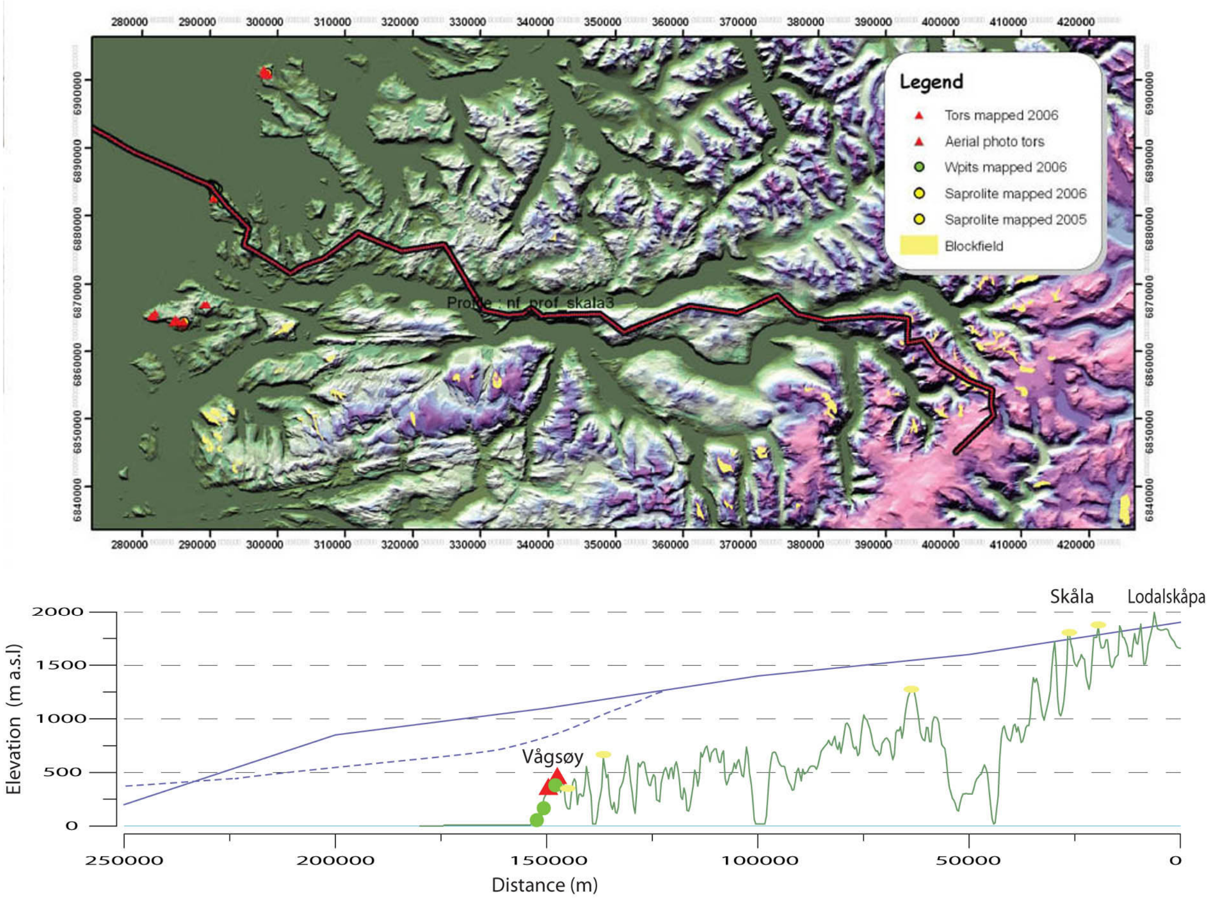

Fig. 28. Mapped relict non-glacial landforms, surfaces and deposits in the Nordfjord area (Fredin, unpublished).

Vågsøy

Vågsøy is an island situated on the coast north of Nordfjord and exhibit a quite considerable relief with the highest summits close to 600 m.asl. and steep coastal cliffs (Fig 29). The relief in “inland” areas of Vågsøy is undulating and summit areas appear to be of significant age and exhibit non-glacial landforms such as tors. However glacial troughs (cirques and valleys) are incised into the topography and numerous small lake basins bear witness of Quaternary glacial erosion.

Vågsøy have probably been overridden by several Quaternary glaciations (Fig. 28; Longva et al., 1983) with main ice flow directions from east (Nordfjord) or from north-east during the last glacial maximum (Nygård, 2003). Fast flowing glacier ice (ice streams?) was situated in throughs and fjords whereas less erosive ice was situated on uplands. The last main deglaciation started at the continental shelf break at about 20 kyr ago and Vågsøy was ice free about 13-14 kyr ago. Glaciers were not present on Vågsøy for a couple of thousand years but was reformed again during Younger Dryas, which will be discussed at the Kråkenes cirque locality.

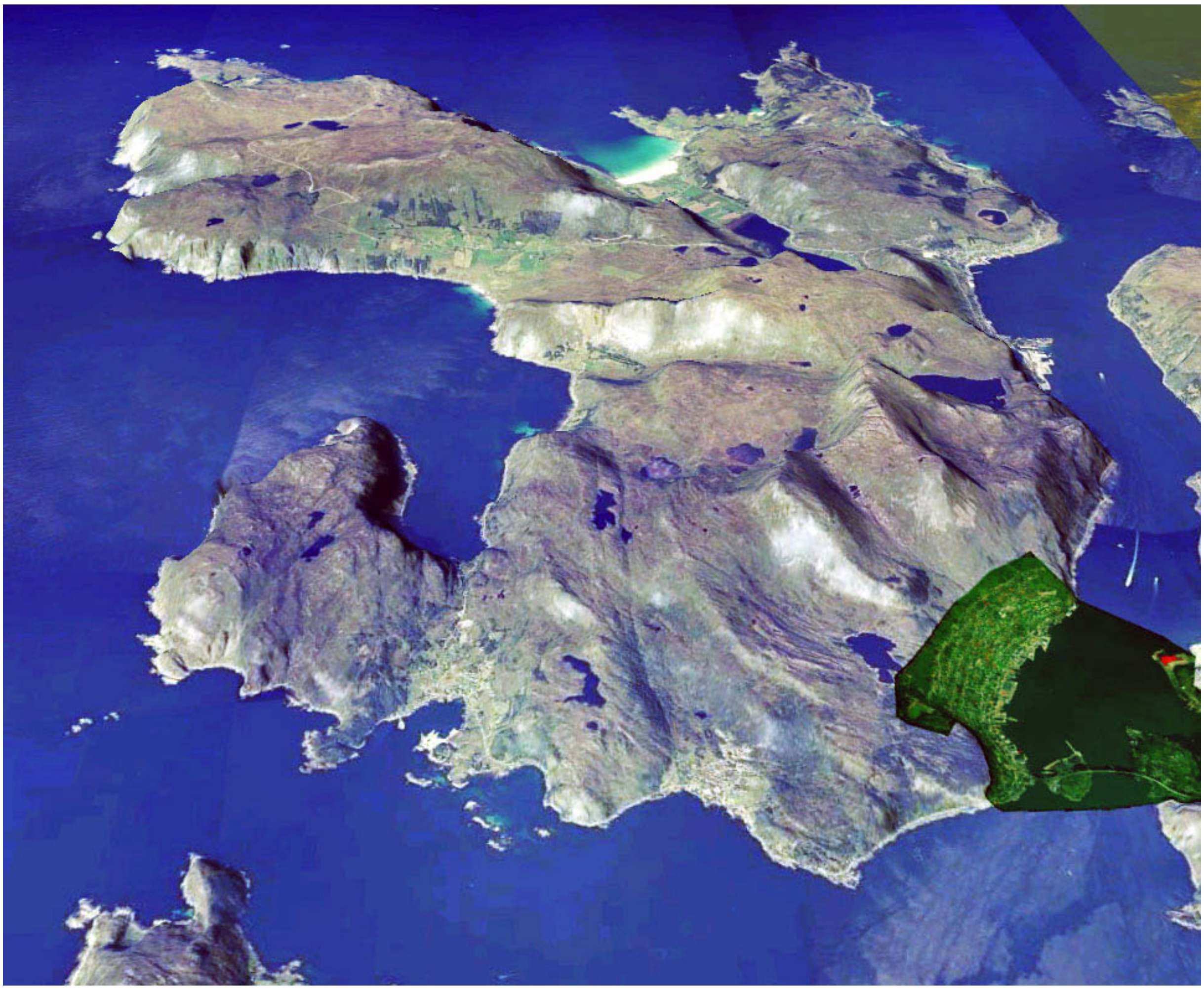

Fig. 29. 3D-view of Vågsøy towards N.

The climate of Vågsøy can be described as oceanic (maritime west coast) with an average annual temperature of 7.1°C (min 2.5°C in January and max 13.0°C in Aug) and precipitation of 1280 mm/yr (min in April and max in Sept.) (http://www.met.no). The westernmost tip of Vågsøy (Kråkenes lighthouse, locality) is considered to be one of the stormiest localities in Norway (which is a stormy place to start with).

Stop No 1: Kannesteinen (61º 58.02' N, 6º 55.13' E)

Kannesteinen is a famous natural sculpture on the west coast of Vågsøy. It is the result of coastal abrasion, where wave action have eroded and cut into a rock platform and leaving a rock pedestal almost defying gravity! The Kannesteinen is definitely a Holocene feature and cannot have survived glacier overriding. It can be noted that outside of the Kannesteinen, there is a narrow "strandflat" developing. The whole Nordfjord coast is surprisingly devoid of a strandflat, whereas prime examples of Norwegian strandflat (50 km wide) are situated on the Møre coast just 150 km north.

Stop No 2: Saprolite at Movatnet gravel pit (62° 00.44' N, 5° 01.20' E) (Fig. 30).

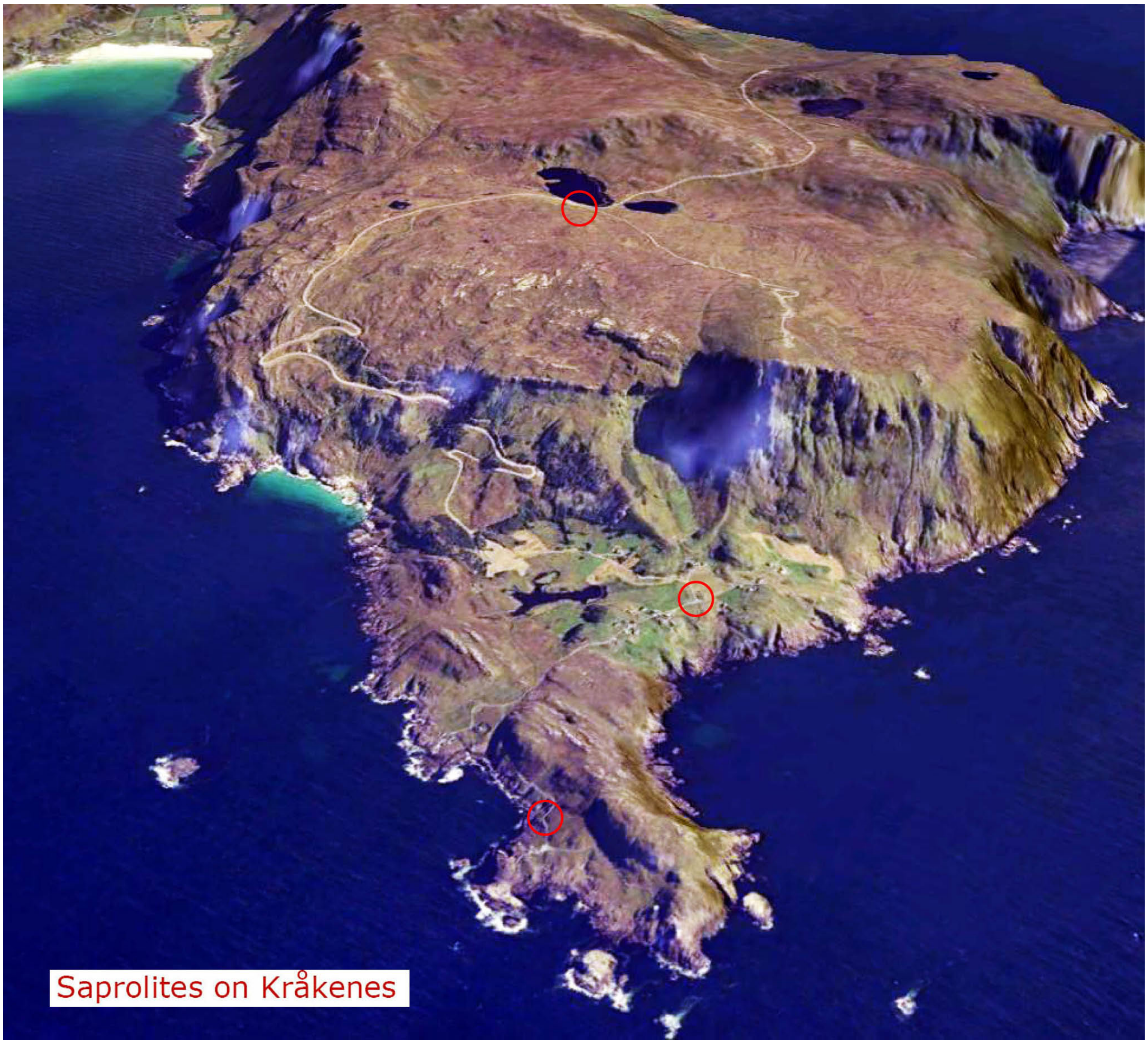

Fig. 30. 3D view of westernmost Vågsøy and Kråkenes towards SE. Mapped saprolite localities are marked with red rings.

This locality was described by Roaldset et al., (1982) and Longva et al., (1983) as a pre-Quaternary saprolite (in-situ weathering). They based their interpretation mainly on morphology of boulders in the gravel pit where e.g. an apparent core boulder is clearly visible today. Moreover, interpretation of XRD spectra from gravel samples indicates presence of the clay mineral kaolinite, which can be used as indicator of chemical weathering at an advanced stage. In addition till samples from the surrounding drift blanket also were interpreted to contain kaolinite, which support the notion that pre-Quaternary saprolites are present in the area and that these saprolites were mixed into the glacial till.
A re-examination of this locality has thrown some new light on these findings (Fredin et al., in prep). The matrix is sandy-grussy and contains very little fines and clays. XRF analysis of main and trace elements have revealed that between 28% and 37% of the parent rock has leached out as chemical weathering thus producing the weathering soil (Brimhall calculations), where Si + Al + Na + K account for about half of the leaching. XRD data suggest certain presence of chlorite (indicative of incipient weathering) and possibly presence of smectite, kaolinite and serpentine. It should however be noted that separation between

chlorite and kaolinite is notoriously difficult (Moore & Reynolds, 1997). A closer look at the eastern part of the pit shows a contact zone between augengneiss and gabbro. It is thus suggested that this locality is not caused by deep weathering but by contact metamorphosis or hydrothermal alterations that has weakened the bedrock and caused leaching of minerals. However, the apparent core boulder in the western part of the pit sure looks like it was formed through deep weathering and cannot be reconciled through the “new” model.

Stop No 3: Kråkenes cirque and moraine (62º 01.62’ N, 5º 0.0’ E)
by Eiliv Larsen[3]

3

At the outermost coast, on the island of Vågsøy, there is a cirque containing a well-developed cirque moraine with a maximum distal height of 16 meters. Melt-water from the glacier that occupied the cirque drained into the small lake Kråkenesvatnet depositing laminated silt and clays. Radiocarbon dates obtained on gyttja silt below and above the glaciolacustrine sediments demonstrated that the cirque glacier formed and disappeared again during the Younger Dryas, i.e. within some 1000-1200 years (Fig. 31). Thus the cirque glacier was not an ice remnant left behind as the ice sheet retreated from the area some 2000 years earlier. This is contrary to cirque glaciers further inland that were left behind as the ice sheet retreated, and only experienced expansion during the Younger Dryas.
Lee-side accumulation of snow by wind and avalanching into the cirque was crucial for growth and to maintain the cirque glacier once summer temperatures were low enough. At maximum, the glacier likely was in equilibrium with climate. The initial retreat from the maximum position might have been triggered by fall-out of volcanic ash from Iceland, but the continued retreat was due to increased ablation season temperatures.
From cores in the lake, seismics on the delta and measurements of the marginal moraines, the sediment volume produced by the cirque glacier was calculated. This was recalculated to bedrock volumes and distributed on the glacier area in order to estimate the erosion rate of the cirque glacier. Given an erosion period of 700 years averaged over the entire cirque area, this indicates an erosion rate of 0.5 to 0.6 mm pr. year. With a constant erosion rate, the cirque could form in some 80.000 to 125.000 years, but obviously cirque erosion has to be distributed over many glaciations.

[3] Geological Survey of Norway

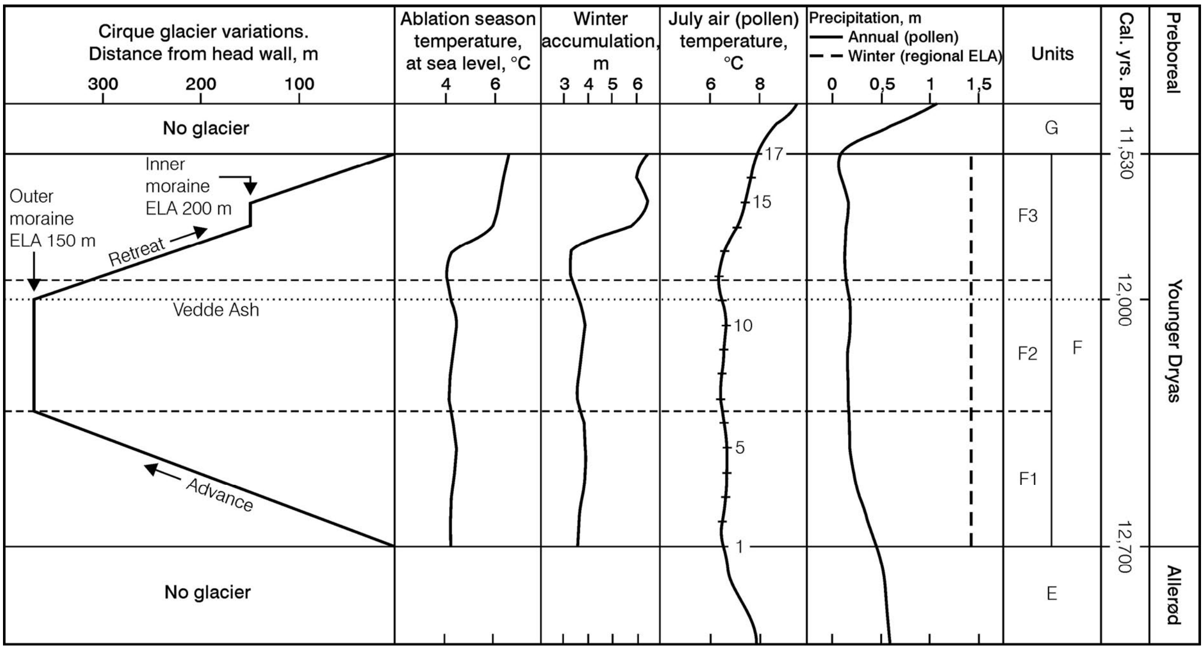

Fig. 31 (Previous page). Time-distance diagram showing inferred cirque glacier variations. July air temperatures and annual precipitation based on terrestrial pollen and spore data are from Birks et al. (2000). Other calculated climate parameters are from Larsen & Stalsberg (2004). The position of the Vedde Ash bed (Mangerud et al. 1984) is dotted. The calendar age scale is according to Gulliksen et al. (1998) and Birks et al. (2000).

Stop No 4: Saprolite in road cut close to Kråkenes light house
(62º 02.08’ N, 4º 59.23’ E)

This locality in Gabbro (Fig. 30) looks like a deep-weathering sandy saprolite with rounded joints, exfoliation and core boulders. It is capped by peat and possibly thin drift cover. The locality has been above the marine limit throughout the Holocene (the marine limit is at about 9 m.asl. during the Tapes transgression). Grain-size, XRD and XRF data is very similar to the Movatnet locality indication about 30% loss of material through leaching and inconclusive clay mineral assemblage.

It can be discussed whether this is a pre-Quaternary saprolite or if it is possible to form such a weathering profile in Gabbro during the Quaternary or even the Holocene. Small weathering pits in the surroundings certainly indicates that the local bedrock is susceptible to weathering in this corrosive environment (more on this on the next locality) but it also seem unlikely that this 4 m profile would evolve through the 14 kyr Holocene history. Maybe the answer is somewhere in-between, since this is a coastal locality it has been ice free during longer time periods than more inland areas (closer to the ice margin) and have had a significant part of the Quaternary period to evolve?

Stop No 5: Weathering pits and tafonis at Kråkenes (62º 02 N, 4º 59’ E)

This is not a single locality but an area with spectacular examples of weathering pits and Tafoni. It will require a bit of walking over sometimes steep and slippery terrain – take care! Already at the previous locality and around the Kråkenes lighthouse there are widespread occurrences weathering pits in the Gabbro. Some of it could probably be classified as honeycomb or alveoli (Bourke and Viles, 2007). Walking up on the ridge south-east of the light-house reveals even more and spectacular weathering pits. There are two distinct classes of weathering pits; classical weathering pits have developed in augengneiss, Fig. 32A and cavernous pits (Tafoni type) have evolved in Gabbro, Fig. 32B. . There are two distinct classes of weathering pits; classical weathering pits have developed in augengneiss, Fig. 32A and cavernous pits (Tafoni type) have evolved in Gabbro, Fig. 32B. Lithology thus has had a key influence on the weathering morphology but also distance to the sea has had significant impact with numbers and magnitude of the weathering pits decreasing away from the sea. It thus seems evident that large tafonis have evolved in Gabbro close to the sea and smaller weathering pits can be found in augengneiss away from the sea, Fig. 33. Some exceptions exist, notably beneath erratics were deep weathering caverns have developed.

A wide spectrum of weathering morphology can be observed at this locality. All stages of cavernous weathering (alveoli, honeycomb, tafoni) can be observed. Many weathering pits are interconnected through channels and the structural control is oftentimes also evident with rills forming along foliations.

Again, the first question to discuss is the age of the weathering pits. Intuitively, the well developed weathering caverns and pits indicate a long formation history. On the other hand, we know that this area has been overridden by the Weichselian ice sheet and recent cosmogenic data indicate that this area was indeed glacially eroded and then deglaciated at

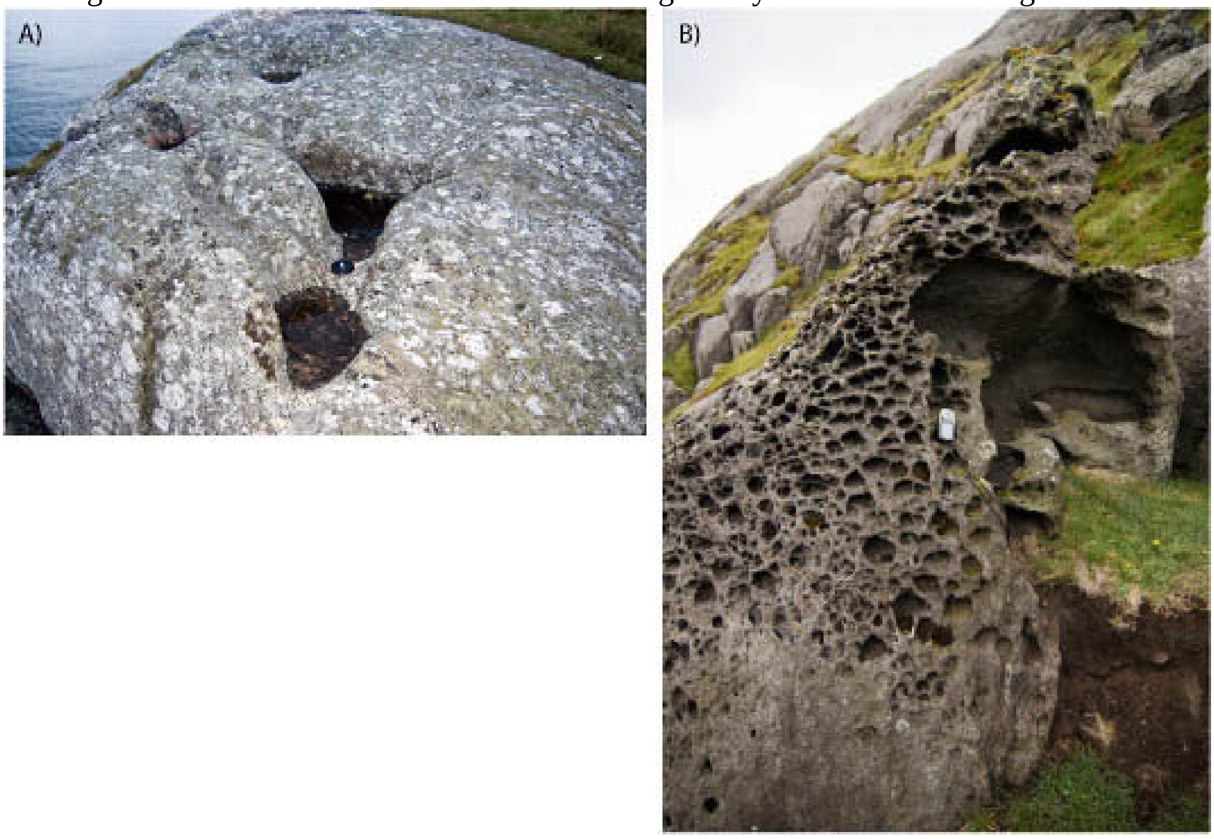

Fig. 32. A) Picture of weathering pits interconnected with channels in Augengneiss. B) Tafoni in gabbro. Photo: Ola Fredin.

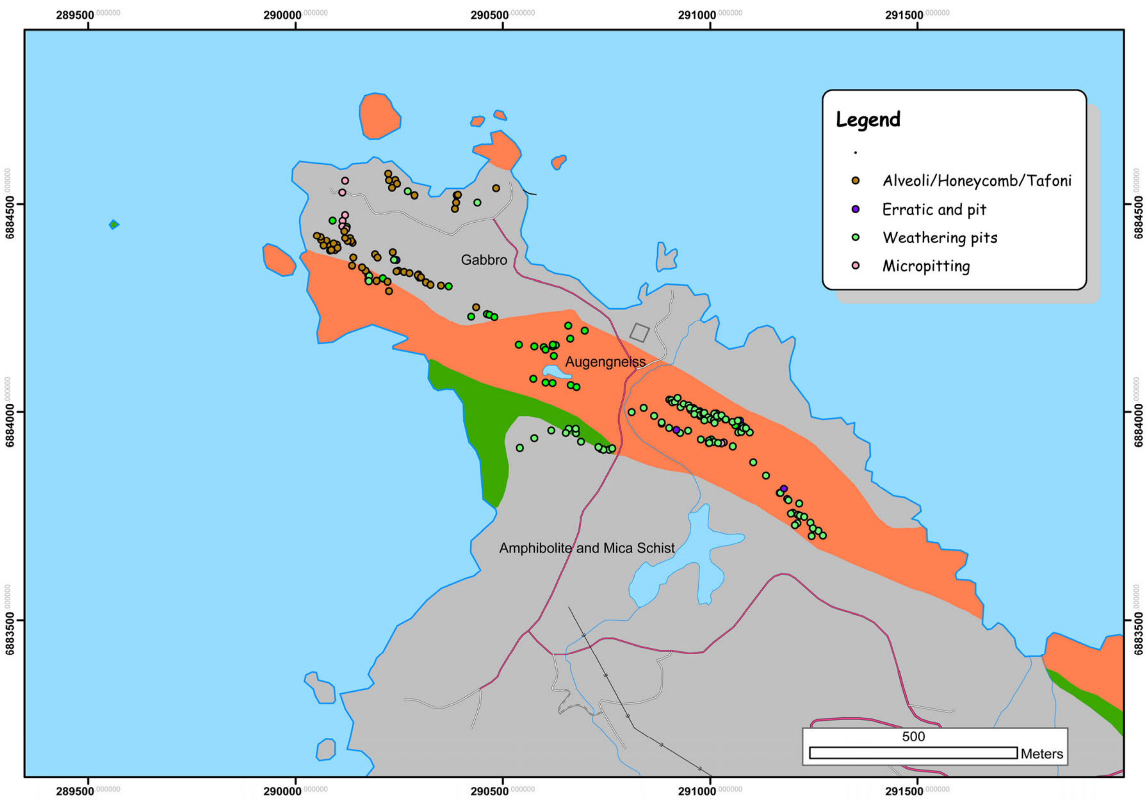

Fig. 33. Bedrock geological map and mapped weathering pits showing the morphological dependence on lithology.

about 14-15 kyr ago. A Holocene age of the weathering pits thus seem likely. Development of weathering pits and in particular tafoni are dependent on chemical attack by salty solutions (Bourke & Viles, 2007), which is a condition amply satisified at Kråkenes. Empirical equations from other parts of the world indicate that it indeed is possible to grow 50 – 100 cm tafonis in about 15 kyr (Norwick & Dexter, 2002). A Holocene age of the observed weathering pits thus seems plausible.

SEDITRANS

SEDITRANS is a project funded by the Norwegian research council dealing with Holocene valley-to-fjord sediment transport and is based in Nordfjord. This area is particularly suitable for this type of study since a wide array of subsystems is present; e.g. glacial processes, slope processes, lacustrine- and marine sedimentation (Fig. 34).

Previous investigations within fjords have resulted in increased understanding of fjord sediment fill (Holtedahl 1975; Syvitski & Shaw 1995; Sejrup et al. 1996; Aarseth 1997). However, relatively little is known about the sediment fill of the sub-aerial valley extensions of the fjords, and even less about the combined fjord-valley systems. Studies of sediment transport in the different parts of the valley-fjord system and sediment accumulations within it, makes it is possible to reconstruct the evolving sediment filling in three dimensions from the present glacier, through the valley, and into the fjord.

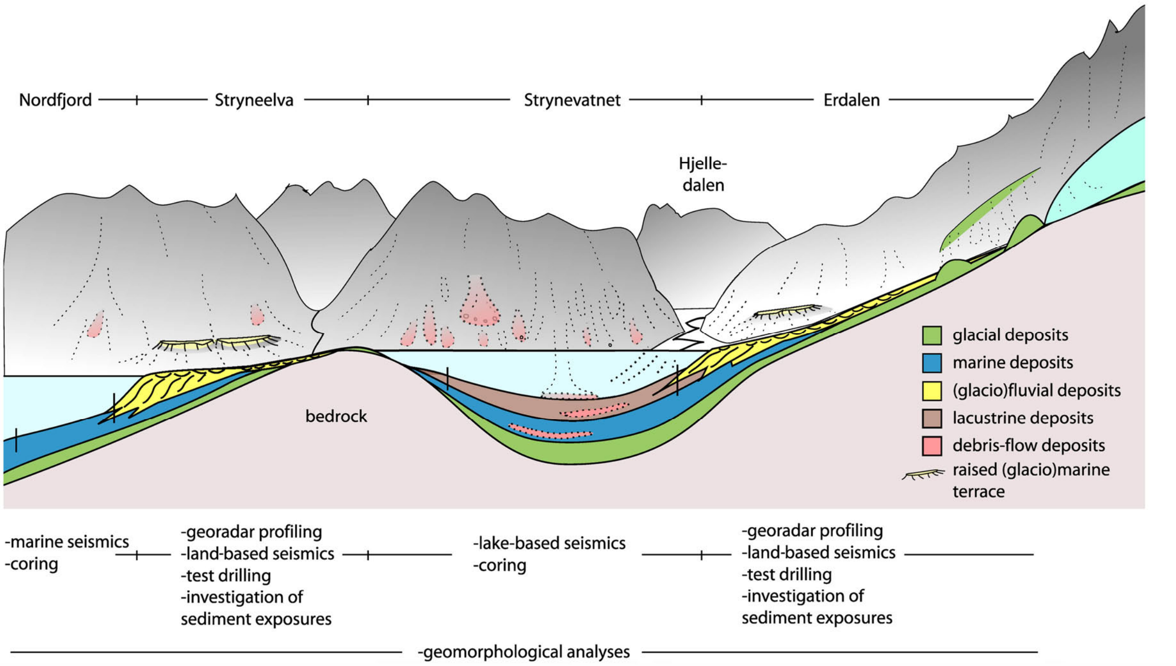

Fig. 34. The subsystems of the SEDITRANS project.

Most of our studies are within (younger than) the Younger Dryas ice margin at Lote/Anda of Nordfjord and concentrate on Holocene processes and sediments. The project is still running (at the time of writing) and only initial data are available, including;

- Detailed bathymetry and seismics (TOPAS) in the marine and lacustrine systems
- Detailed maps and geophysics of terrestrial deposits and landforms, including updated deglaciation mapping and chronology
- Past and present sediment budgets in the Bødalen and Erdalen sub-catchments

At the time of writing (Jan-2008) modeling of sediment volumes and source-to-sink processes has started and hopefully results will be available for the excursion in august 2008.

Day 4: Sunday 17th August: Nordfjordeid – Stranda – Valldal - Geiranger

Introduction

On this day we will demonstrate the geology of valleys with thick glacial accumulations as well as the contrast to this: fjords, lakes and valleys with extreme glacial erosion. The extreme relief has triggered prehistoric as well as relatively resent mass movements and tsunamis.

Before the glacial re-advance in the Late Younger Dryas the sea inundated the area now occupied by Lake Hornindalsvatn and marine shells, dated to Allerød (11 360 +/- 70 B.P.), are found at the eastern end of the lake (Fig. 35). During the re-advance the glacier advanced to Nor where an outwash delta was formed. The glacier dammed the upper Hornindal valley and a lateral shoreline from the lake can be seen. The drainage during this phase was towards NE, over the pass (389 m a.s.l.) and down to Hellesylt where we will have our packed lunches.

From a rest area just north of Hellesylt we can see the large potential Åkneset slide area 6 km further to the NE. The unstable part of this mountain slope comprises 10-15 mill m^3, but volumes of 35 m^3 cannot be excluded. This could cause a tsunami that may reach a height of 35 m in the community of Hellesylt.

At Stranda we will get detailed information about the Åknes-Tafjord early-warning center After a short ferry crossing we will board a small ferry at Valldal for a 2 h 15 min cruise on the Norddalsfjord, Synnulvsfjord and the UNESCO heritage site Geirangerfjord. The ferry passes just below the Åknes rockslide before we arrive in Geiranger relatively late (19:00) in the evening.

The night will be spent at the Grande Hotel situated at the foot of “The Eagle Road”, 2 km NW of Geiranger village. (“Grande” is a Norwegian name meaning a river delta, here represented by an alluvial fan built out into the fjord).

Eidsdalen

The relatively broad Eidsdalen valley between the Eidsfjorden and the Lake Hornindalsvatn has a valley fill of glaciofluvial terraces, glaciomarine clays and fine-grained fluvial sediments. Two finds of molluscs in till here yielded ages of 10750 +/-140 BP and 10930 +/-160 BP (dates given as radiocarbon yrs BP), predating the Nor moraines at the outlet from Lake Hornindalsvatnet (Fareth 1987).

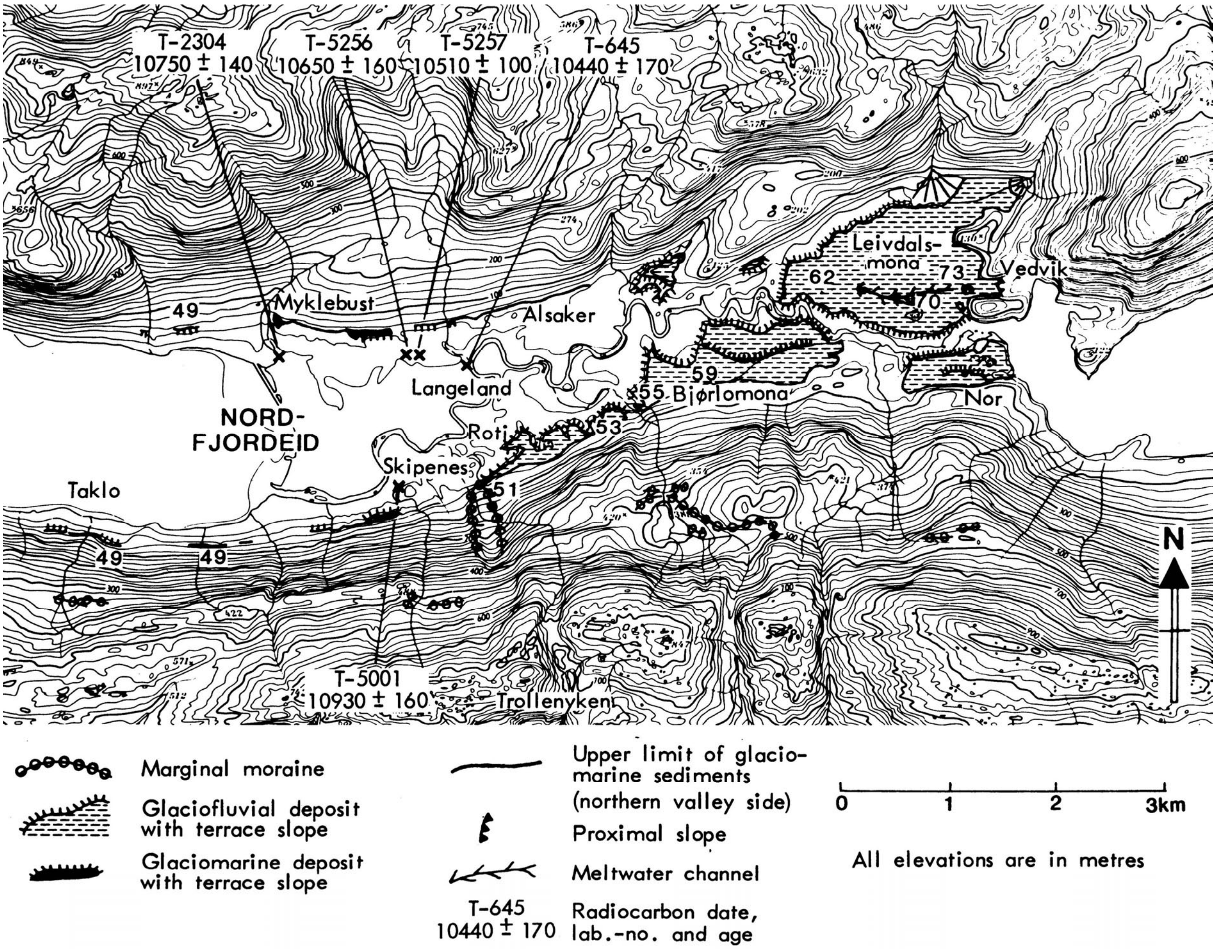

Fig. 35. Map of the Eid valley, Nordfjordeid with marginal moraines, terraces and radiocarbon dates of marine molluscs (Fareth 1987).

Stop No 1: Nor sand and gravel pit, Vedvikmona (61° 55.00' N, 6° 05.73' E)
Three dates of molluscs from the glaciomarine deposits distal to the Nor moraines range from 10440 +/-170 to 10650 +/-160 BP. This confirms the early age estimate that the ice-front deposits at Nor are of Younger Dryas age. Although these deposit are called the Nor moraines, no real moraine ridges are found at the type locality. Instead a large outwash (sandur) deposit is built up to maximum of 73 m a.s.l. It slopes towards the west and stretches for 3 km along the southern valley side, (Fig. 35.). The upper marine limit is at 49-55 m a.s.l. in this valley, lowest to the west. The Eidselva river has cut down through terraces during the Holocene. It has now incised meanders in the upper part and a series of meanders on the floodplain in the lower part. The fluvial sediments have a general thickness of 2-3 m above the eroded glaciomarine silty clays (Klakegg & Nordahl-Olsen 1985).

Stop No 2: Lake Hornindalsvatn (514 m deep!)

This 24 km long lake is the deepest lake in Europe and resembles a long fjord. Before glacier re-advance in Younger Dryas it was a real fjord and marine molluscs found near the eastern end of the lake have been dated to the Allerød interstadial (11 360 +/- 70 BP). During the Younger Dryas Stadial this northernmost glacier tongue in the Nordfjord area advanced and filled the present lake basin completely. A glacier-dammed lake was formed in the upper Hornindal valley between the glacier and the present watershed to the northeast. A 4 km long and 10-60 m wide lateral shoreline was then eroded into the till on the south-facing valley slope. The terrace slopes toward the pass point at 389 m a.s.l. (Fareth 1987). Fig. 36 shows three phases of the Hornindal lake: glacial, marine and present lake phase.

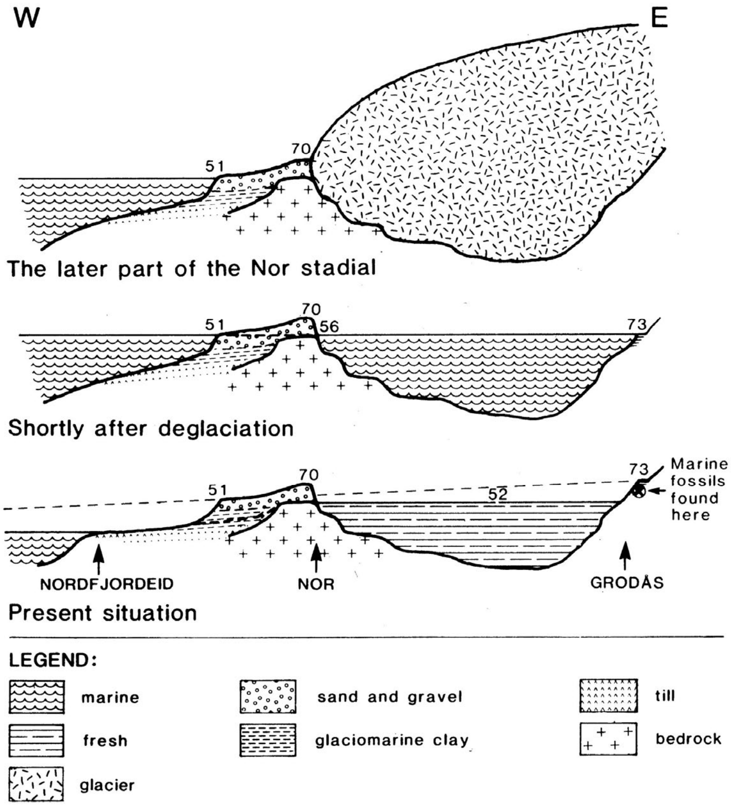

Fig. 36. Phases of development in Lake Hornindalsvatnet during the deglaciation (Fareth 1987)

Seljeset:Terminus of Y. D. glacier tounge in the Hornindal valley

Stop No 3: Lyngvollen: Shoreline in glacier dammed lake (61º 59.92’ N, 06º 36.61’ E)

During the Nor stadial the glacier-dammed lake drained over the present pass point (389 m a.s.l.) and down to Hellesylt. The corresponding lake level is marked by prominent shorelines and terraces. A stop will be made near the present watershed, at Lyngvollen. Further west a terrace can be traced for 4 km on the northern side of the valley. The width varies from 10 to more than 60 m. During Younger Dryas many local glaciers of the mountain areas on both sides of Hornindal valley extended far down towards the main valley, thus causing a considerable increase in the sediment supply. The glacial lake level is rather conspicuous in the field at Lyngvollen where a continuous, in places more than 100 m wide terrace extends from the lateral deposit towards Northeast at an altitude corresponding to the present watershed (Fareth 1987).

Stop No 4: Hellesylt. Packed lunch (62º 05.16’ N, 06º 52.18’ E)

The village Hellesylt is the closest inhabitated area to the Åknes rockslide. A large collapse and rockslide from Åknes can initiate damaging tsunamis towards the settlement in Hellesylt. Tsunami modelling indicate run-up heights of up to 35 meters. Early-warning and evacuation plans are today operative for the area.

Stop No 5: Ljønibba view point, Photo stop (62º 07.74’ N, 06º 55.43 ’ E)

From here we are able to see the Åknes rockslide in profile. Using binoculars one can see some of the installations and helicopter pads on the upper part of the mountain slope.

Stop No 6: Stranda (62º 18.50’ N, 6º 56.90’ E)

An early-warning center has been established in the municipality of Stranda in order to handle all issues related to the monitoring and early-warning. Geoscientific leader Lars Harald Blikra will give an orientation at the activity and organization of the center (Fig. 37).

Ferry Stranda – Liabygda

Short ferry crossing of the Synnulvsfjord. 15:15 – 15:30.

Tourist Ferry Valldal – Geiranger Valldal (62º 17.90’ N, 7º 15.80’ E)

16:45 – 19:00.

Just after we leave the Valldal ferry quay we pass above the submarine part of the Younger Dryas terminal moraine in the Norddalsfjord (Fig. 38). Here the moraine ridge itself does not constitute a ridge on the bottom of the fjord because it has acted as a sediment trap for younger glaciomarine sediments as well as Holocene prodelta sediments from the Valldalen river. Distal to the moraine there are sequences of acoustically laminated glaciomarine sediments more than 150 m thick. Younger slumped material is also present on the gently sloping fjord bottom (Sejrup et al. 1996).

Åknes rockslide (62º 10.50’ N, 7º 0.2’ E)

The Åknes rockslide is situated within the Western Gneiss Region, dominated by gneiss of Proterozoic age. The morphological investigations shows several characteristic features, listed below (Fig. 37). North of the Åknes rockslide, we will have a view towards a large slide scar on the western side of the fjord. Bathymetrical and seismic data demonstrates that gigantic rock avalanches have been released here, filling up large parts of the fjord basin. The rockslide at Åknes is located on the western side of Sunnylvsfjorden, stretching from about 900 m at the upper tension cracks to about 100 m above sea level at the base. Geological and geophysical investigations at Åknes indicates that the unstable area covers almost 0.8 km^2.The Åknes rockslide is a large rockslide of possible 30-40 million m^3, moving with a velocity of 3-10 cm/year. The monitoring systems is today based on extensometers, single lasers, GPS, total station, geophones, climate station and borehole instrumentation (inclinometers and piezometers). The data is implemented in an integrated web-based system. Major challenges are linked to the steep terrain, remote setting and problems with rockfalls and snow avalanches. Major effort has been put on to get reliable operational power and communications systems. The movement data so far demonstrates a continuous movement

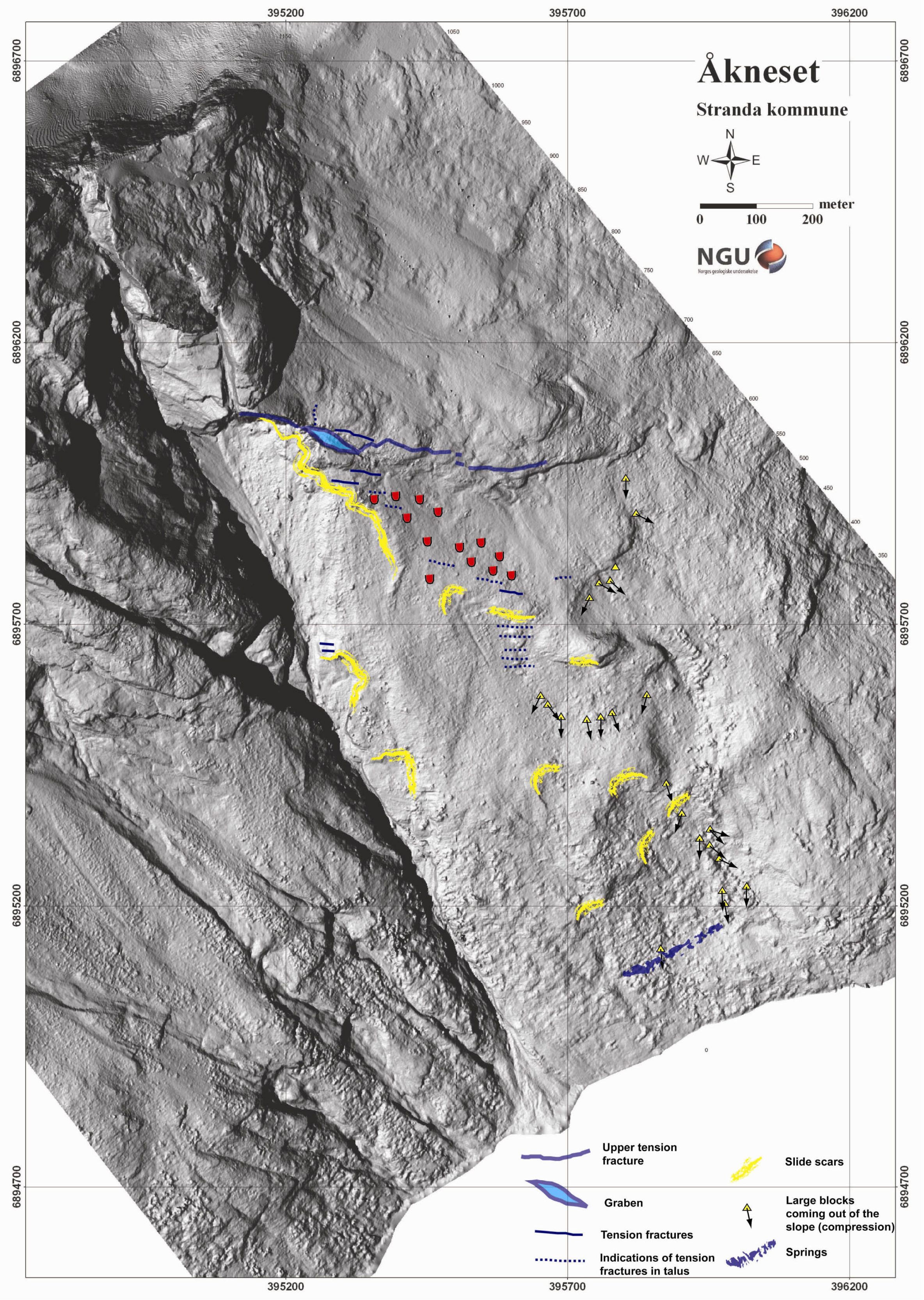

Fig. 37.Shaded-relief map of the Åknes rockslide, with the morphological characteristics indicated.

1. An about 500m more or less continuous back crack (Upper tension fracture).

2. A large depression in the upper western corner of the rockslide, developed into a graben structure. The total vertical displacement is from 20-30 m.

3. A series of tension fractures from the upper to the middle part of the slope. They are oriented in a WNW to ESE direction and shows up to 1 m openings in bedrock and depressions and collapse structures in the blocky colluvial debris.
4. Prominent slide scars along the southwestern canyon. Historical data indicates a slide in the upper part in the late 1800, and slides also in 1940 and 1960.
5. Small slide scars in the lower part of the rockslide.
6. Large blocks or parts of the rocks is coming out of the slope at two particular areas, one in the middle part and one area in the lowermost part.
7. Distinct water springs at the lowermost part of the slope at about 100masl.

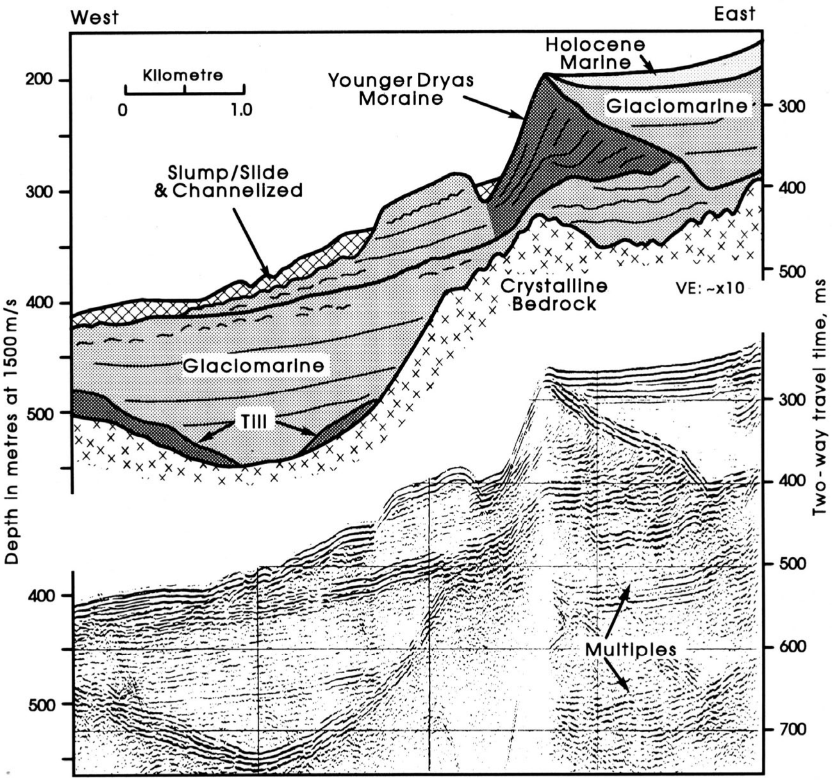

Fig. 38. Interpreted seismic profile of the Younger Dryas terminal moraine and distal sediments in the Norddalsfjord (Sejrup et al. 1996).

during the entire year, but with significant seasonal changes. Based on the historical data from the Åknes rockslide and information from historical rockslide events elsewhere, preliminary early-warning levels have been proposed (green, blue, yellow, orange and red levels)

Fjord intersection Synnulvsfjord – Geirangerfjord.

At the intersection between the Synnulvsfjord and the Geirangerfjord the two fjord glaciers nearly coalesced during the Younger Dryas ice advance. In the Synnulvsfjord a sharp moraine ridge with foreset beds and a sharp crest at 140 m water depth is found just outside Bjørkeneset. In the Geirangerfjord the terminal moraine is more complex. Strong reflectors above the acoustic basement at both localities are interpreted as older basement till, whereas the moraine in the Geirangerfjord area also has younger rock slides capping the foreset beds (Fig. 39).

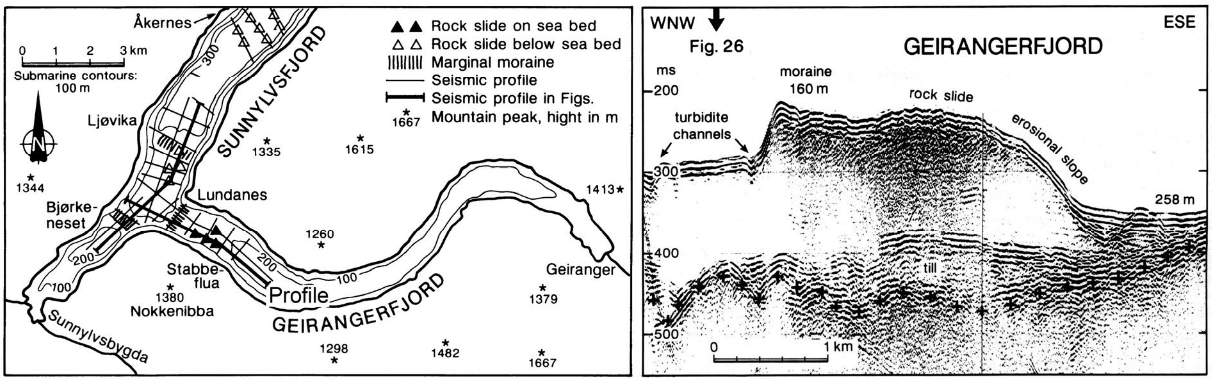

Fig. 39. Map of the inner part of the Synnulvsfjord-Geirangerfjord area (left). Seismic profile (Sparker) of the Younger Dryas terminal moraine in the Geirangerfjord (right).(Aarseth et al. 1997).

Former settlements along the Geirangerfjord.

From the ferry we can observe old farmhouses, either on small fan-deltas along the fjord or on valley shoulders where the topography enabled houses to be built. Some of these farms were inhabited up to World War II. Many of them have been restored by the society “The Friends of Storfjord”. The nomination document for the World Heritage Site lists 19 abandoned farms along the fjord. The buildings were placed in sheltered areas to prevent damage from rock falls, floods or snow avalanches. A Fjord Centre at Geiranger demonstrates the struggle for existence at these farms.

<u>Evening meal and stay overnight at Grande Hotel,</u> 2 km north of Geiranger village.
(+47) 70 26 94 90 (62º 6.98’ N, 7º 11.11’E)

Day 5: Monday August 18th Geiranger – Briksdalen – Loen -Stryn

Introduction

This morning will bring us to three of the best viewpoints in Geiranger (“The Eagle Bend”, Flydalsgjuvet canyon and Dalsnibba mountaintop at 1476 m a.s.l. (if weather permits).
The first and the second view will demonstrate the fjord morphology with a winding fjord, truncated spurs and valley shoulders. Several of the latter have abolished farmsteads.
From the mountaintop we hopefully will be able to sea both the fjord deep down and the alpine topography of this part of the county “Møre and Romsdal”. Down on the main road again (R-63) we pass Langevatnet were a steep mountain slope just above the road is susceptible to snow avalanches during springtime. This road is closed in wintertime and ice may stay on the lake till early August.

The relatively new road to Stryn (R-15) leads through three long tunnels. In the Grasdalen valley there is a snow avalanche research station run by Norwegian Geotechnical Institute. Large debris cones are built to protect the road against avalanches. The Stryn valley is a classic U-shaped valley and a photo stop will be made at the intersection to the old road just outside the third tunnel.

Briksdalsbreen glacier has long been a tourist magnet, but also the subject for glacial geologic research as this short outlet glacier from the Jostedalsbreen glacier has a very short (3-4 years) responding time to the glacier budget driven by summer temperatures and winter precipitation.

Lake Lovatnet has been the site of several large historic tsunamis generated by rockfalls from the mountain Ramnefjellet. Several villages were wiped out and 135 people were killed in the two largest accidents (1905 and 1936).

Stop No 1: "The Eagle bend". View of the Geirangerfjord (62º 07.57' N, 7º 10.14' E)

After admiring the view and photographing the fjord (Fig. 40), there is time to look at the geology! The foliated gneissic rock at "The Eagle Bend" has pronounced sheeting with gentler dip towards the fjord than the hillside. These so called "valley-joints" have certainly played an important role in the denudation processes. They have enabled glacial plucking during glacials and stadials as well as increased lateral mass movements along the steep valley sides during ice-free periods. The interaction of the oblique sheeting and the vertical joint system is clearly producing potential rock falls. For construction of a platform like this, one has to take the sheeting into consideration and anchor the platform deep into the rocks. Like the view of the Aurlandfjord from the "Skywalk" at Stegastein one can try to reconstruct valley generations. As in the Nærøyfjord and Aurlandfjord, some of the ledges have been used as foundations for settlers. The waterfall "Seven Sisters" at the distance is falling off the vertical rock wall of a truncated spur.

Fig. 40. View of the Geirangerfjord from the platform at "The Eagle Bend". Photo: I. Aarseth 2007).

Stop No 2: Flydalsjuvet canyon. View of Geiranger (62º 05.50’ N, 7º 13.40’ E)

The second stop at Flydalsjuvet shows in addition to a breathtaking view of Geiranger also the fluvial down-cutting of the river trying to eliminate a glacial step in the valley. The bedrock structures are mirrored in overhanging rock formations where people with no aversions towards giddiness can pose on the edge. On both localities the road authorities have used landscape architects to plan and build the platforms.

Stop No 3: Dalsnibba mountaintop. (1476 m a.s.l.) **(if weather permits)** (62º 03.10’ N, 7º 16.25’ E)

In addition to a distant view of the fjord and the Geiranger village we can see the skyline of “The Sunnmøre Alps”. But even in this alpine landscape the mountain tops have more or less the same altitude: 1600 – 1800 m a.s.l. Very few mountains are flat toped like further south in western Norway. The summit at Dalsnibba has not developed a block field.

Stop No 4: Fonnbu field station for snow and avalanch research, Grasdalen valley (61° 59.85’ N, 7° 18.80’ E)

The Strynefjell mountain road used to be closed in wintertime. The old road is now open in summertime and a summer skiing centre (down hill) has lifts on north-facing slopes. During planning of the new road research was carried out to prevent snow avalanches on the road. A snow avalanche research station has been operated by the Norwegian Geotechnical Institute since 1973 (www.ngi.no). A new modern station was opened in 2006 after a fire. Here we will get information on the research at the station. All kinds of weather diagrams can be found on their internet address: www.fonnbu.no In addition www.snoskred.no gives information on snow avalanche (in Norwegian).

Stop No 5: Videseter intersection R 15 – R 258. View point: Photostop. View of Stryn valley (61° 56.30’ N, 7° 15.75’ E)

Stop No 6: Briksdalsbreen mountain lodge (61° 39.90’ N, 6° 51.25’ E)

After lunch at Briksdalsbreen mountain lodge we will walk towards the glacier (30 minutes, 200 m elevation).

Frontal fluctuations of Briksdalsbreen, a western outlet glacier from Jostedalsbreen in western Norway

by Atle Nesje

Introduction

Briksdalsbreen (11.94 km^2) is a steep outlet glacier from Jostedalsbreen (487 km^2), the largest icecap on mainland Europe. The glacier ranges in altitude from 1910 to 350 m over a distance of 6 km (Østrem et al. 1988). Briksdalsbreen attained its maximum ’Little Ice Age’ position around AD 1760-65 (Pedersen 1976). Annual frontal measurements were started at Briksdalsbreen in 1900 by Johan Rekstad at Bergen Museum. Fig. 41 shows pictures of the glacier at 1871, 1900, 1953 and 1963. In the first part of the 20th century the glacier front was

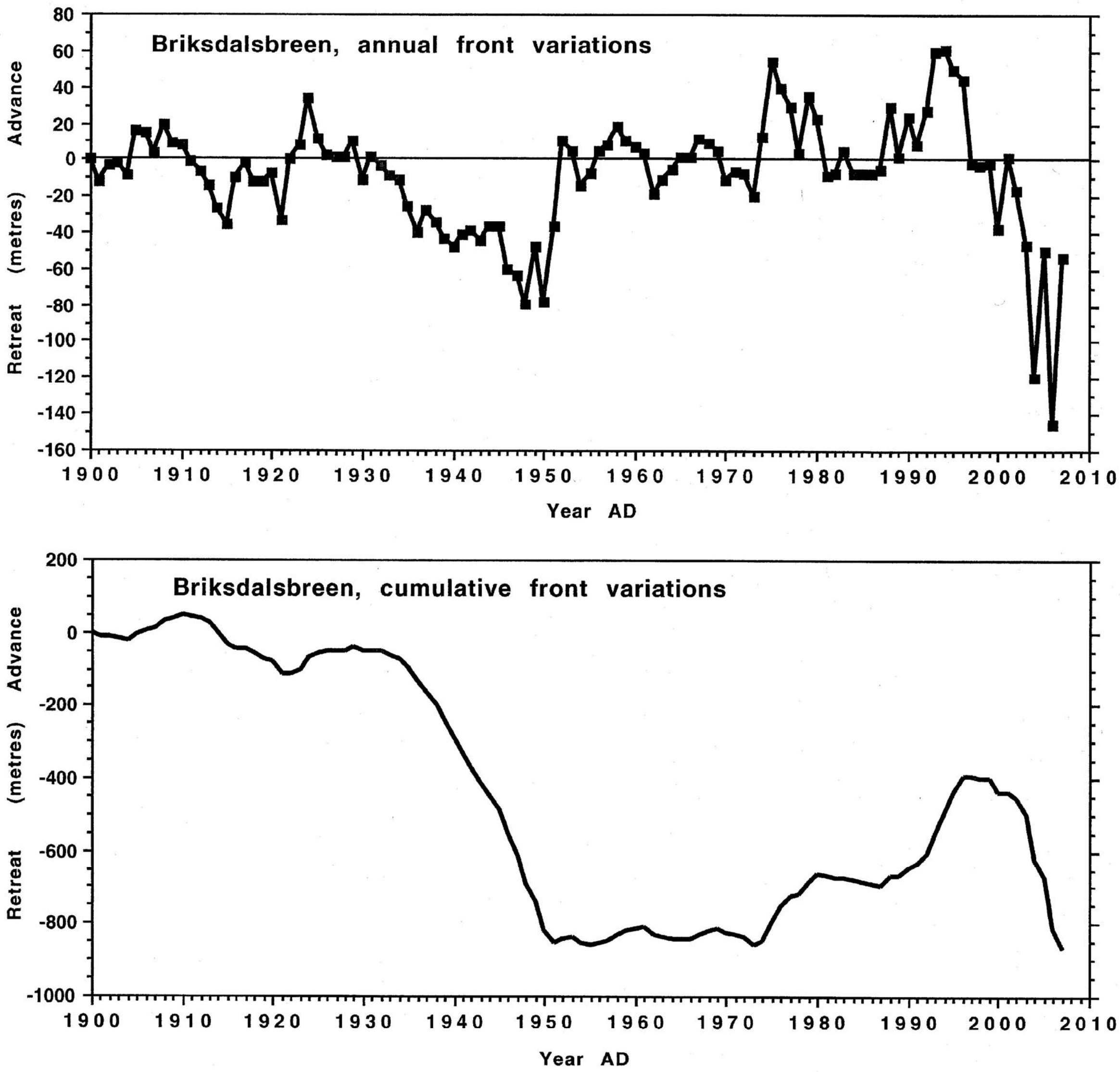

Fig. 41. Upper panel: Annual front variations of Briksdalsbreen AD 1900-2007 (data: NVE). Lower panel: Cumulative front variations of Briksdalsbreen AD 1900-2007 (data: NVE).

in a rather stable position, however, with a minor glacier advance that culminated in 1910 (Fig. 42). During the 1930s and 1940s, however, the glacier front retreated significantly, reaching a maximum annual retreat in 1948 with 79 m. The distal part of the proglacial lake Briksdalsbrevatnet (maximum water depth in the 1980s of 20 m, MacManus & Duck 1988) was deglaciated in the early 1940s, whereas the minimum glacier extent was reached in 1955 (-862 m relative to the 1900 frontal position, Fig. 41). Between 1952 and 1973 the glacier front was more-or-less in the same position, however, between 1974 and 1980 the glacier front advanced 186 m. In 1988 a significant glacier advance started, which culminated in 1994 with 61 m, the largest annual advance recorded in the 20th century. The glacier front reached its maximum extent in 1996, when the glacier tongue covered the entire Lake Briksdalsbrevatnet. After 2001, the glacier front has retreated significantly (maximum annual retreat of 145 m in 2006) and the distance between the 1996 and 2007 frontal positions is 475 m (Fig. 43).

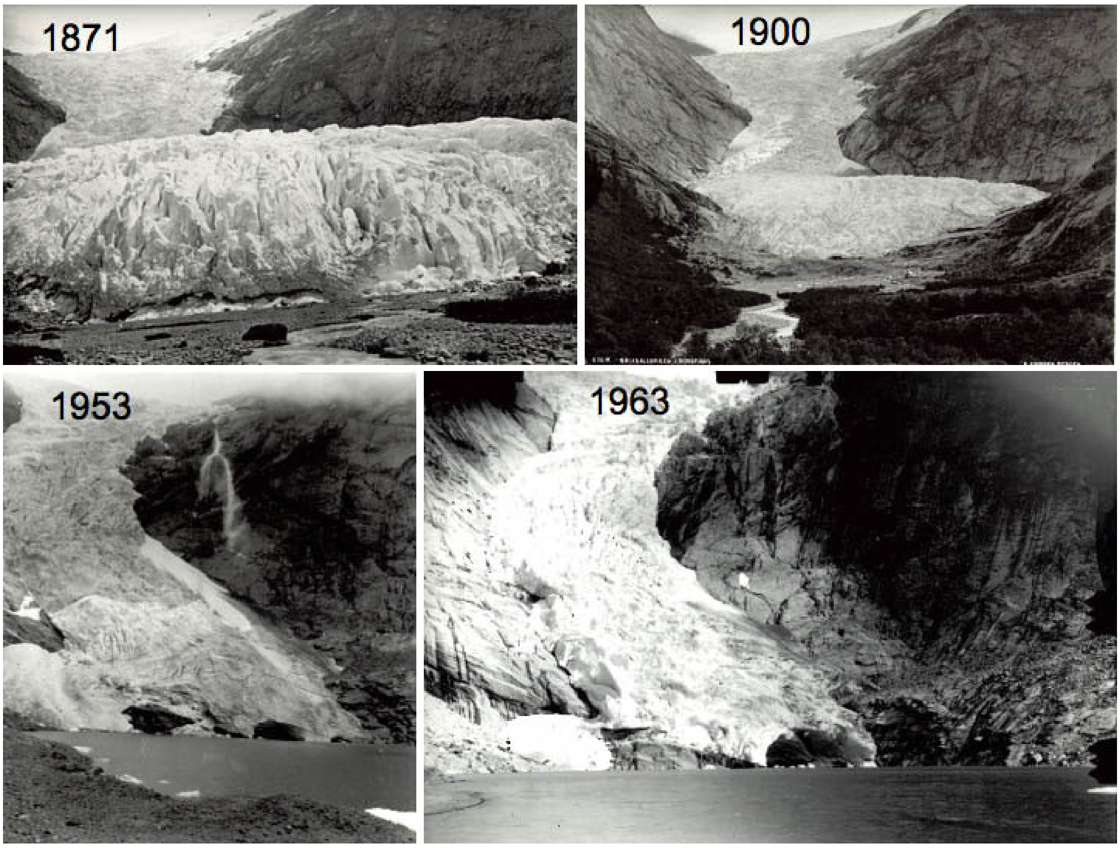

Fig. 42. Briksdalsbreen in 1871 (photo: Knud Knudsen, Billedsamlingen, University of Bergen), 1900 (photo: Knud Knudsen, Billedsamlingen, University of Bergen), 1953 (photo: Olav Liestøl), and 1963 (photo: Olav Liestøl).

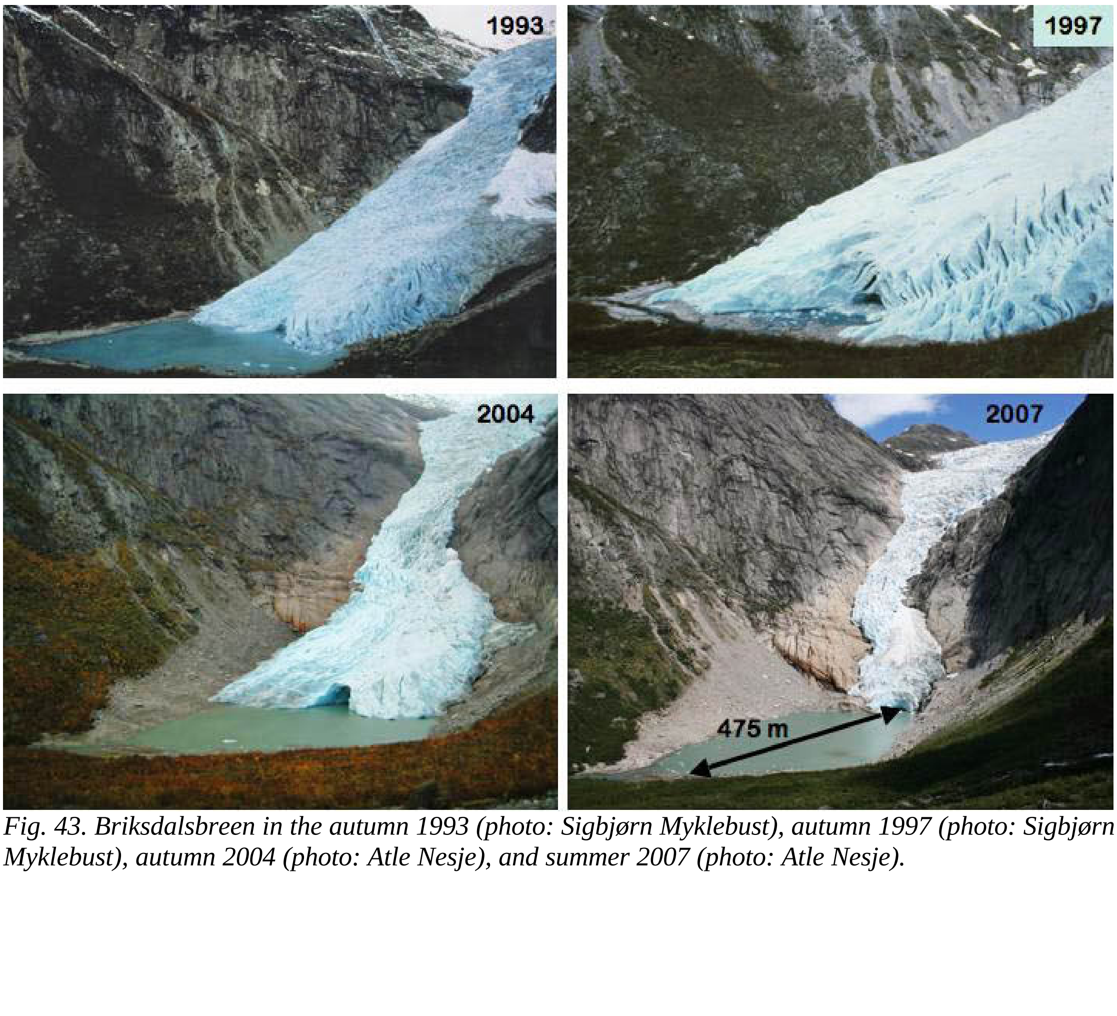

Fig. 43. Briksdalsbreen in the autumn 1993 (photo: Sigbjørn Myklebust), autumn 1997 (photo: Sigbjørn Myklebust), autumn 2004 (photo: Atle Nesje), and summer 2007 (photo: Atle Nesje).

Climate variability in western Norway and terminal response of Briksdalsbreen

Annual glacier-front variations of Briksdalsbreen between 1900 and 2007 have been compared with winter precipitation and summer temperature records from Bergen over the same period in order to evaluate the reasons for the observed glacier-front variations (Nesje 2005). Between 1901 and 1910 the glacier front advanced mainly as a result of lower summer temperatures. Subsequently, the glacier front retreated until 1921 followed by minor advance until 1931. In the period 1932-1955 the glacier front receded as much as 809 m. The significant retreat was mainly a response to high summer temperatures and low winter precipitation, especially during the early 1940s. In 1956 the glacier started to advance as a combined effect of lower summer temperatures and higher winter precipitation. Between 1956 and 1992, the glacier advanced due to lower summer temperatures and higher winter precipitation. In the four years between 1992/93 and 1995/96 the glacier front of Briksdalsbreen advanced 242 m due to high winter precipitation. Between 1997 and 2007 the glacier front receded 475 m, with the largest annual retreat in 2005/06 of 145 m (mean ~40 cm day^{-1}). This is the largest annual recession since the annual frontal measurements started in 1900. The front of Briksdalsbreen was in the autumn of 2007 eight metres behind the 1955 frontal position. The main cause of the significant glacier retreat during the late 20th/early 21st centuries was a combined effect of low winter precipitation and high summer temperatures. The summer temperatures during this period were on the average higher than during the summers in the 1930s and 1940s.

The Briksdalsbreen 1900-2007 glacier record demonstrates that glacier variations not only are a response to ablation-season (summer) temperature, but are also highly dependent on accumulation-season (winter) precipitation.

Dated tree logs

In the 1990s, the glacier front pushed up glaciolacustrine sediments (Winkler & Nesje 1999). In 1995 a well-preserved log was found protruding from the terminal push moraine that was under formation in front of Briksdalsbreen (Nesje 2005). The log, which contained 62 annual rings, was identified as a willow (*Salix cuprea*). The log was dated at the Trondheim Dating Laboratory to be about 8400 calendar years old. Later, two more, well preserved *Salix* logs were found and radiocarbon dated (all three ages are shown in Table 1). The *Salix* logs date from the period immediately before or during the 'Finse event'/'8200 cal. yr BP event' (Nesje et al. 2001). The radiocarbon dates record when the trees died. The three logs may therefore have been buried rapidly in the sediments in the Briksdalsbrevatnet basin before or during the glacier advance associated with this event. The site was overrun by Briksdalsbreen during the 'Little Ice Age' without eroding/destroying the wood remains.

Table 1

Sample no.	Lab. no.	Sample weight (g)	^{14}C age	Intercept cal. age	Cal. time interval	• ^{13}C
Briks-1	T-12175	5.0	7650±85	8405	8485-8340	-26.1
Briks-2	T-12883	7.7	7530±100	8325	8400-8165	-26.1
Briks-3	Beta-158754	-	7340±60	8160	8190-8050	-25.0

Radiocarbon dates of willow (Salix sp.) tree logs found in the frontal moraine of Briksdalsbreen during the advance in the 1990s. The ages and sigma. All samples are dated by the 'conventional' method at the Trondheim

Dating Laboratory, Norway (T) and at Beta Analytic Inc., Florida (Beta). INTCAL98 (Stuiver et al. 1998) was used for calendar age calibration.cal. time intervals are given ± 1

Stop No 7: Bødal, Loen: Rockfalls from Ramnefjellet in 1905 and 1936 and the devastating tsunami waves they generated in Lake Lovatnet

(61° 47.74‘ N, 7° 0.97‘ E)

January 15th 1905 and September 13th 1936 two of the largest natural disasters in modern times occurred in Norway. The rockfalls in 1905 and 1936 triggered large tsunami waves along the shores of Lake Lovatnet. In 1905 and 1936 there were 61 and 74 casualties, respectively, altogether 135. In total, seven rockfalls occurred in Ramnefjellet during the period 1905-1950 (Table 1). It has been estimated that a total of ca. 3 million m^3 of rock have fell down from Ramnefjellet since 1905.

The 15 January 1905 rockfall

Sunday evening 15 January 1905, between 23.00 and 24.00 PM, two loud bangs were heard from Ramnefjellet. A volume of 50,000 m^3 of rock fell down from an altitude of about 500 m and triggered about 300,000 m^3 of till and talus material at the base of the slope. It has been estimated that 870.000 tons of material moved into the lake.
The rockfall generated large waves with a maximum height of 40.5 m above the lake level in Lovatnet (52 m a.s.l.).

The 13 September 1936 rockfall

The September 13 rockfall occurred at 04.30 AM. 1 million m^3 of rock fell down from Ramnefjellet and into the lake. The rockfall generated waves with a maximum height of 74.2 m (Jørstad 1968). The boat ‘Lodalen’ that was washed ashore in the 1905 tsunami was now lifted 33 m higher up to a position 350 m from the shore (the boat can still be seen from the road).

Table 2

Date	Volume rock (m^3)	Volume till/ talus (m^3)	Fallout (m a.s.l.)	Maximum wave height (m)	No. of casualties
15.01.1905	50,000	300,000	500	40.5	61
20.09.1905	ca. 15,000	ca. 50,000	400	>15	0
13.09.1936	1 million	-	800	74.2	74
21.09.1936	ca. 100,000	-	800	ca. 40	0
06.10.1936	?	-	800	?	0
11.11.1936	>1 million	-	800	>74	0
22.06.1950	ca. 1 million	-	800	ca. 15	0

Rockfalls from the mountain Ramnefjellet in Loen 1905-1950.

Submarine slide in the Innvikfjord 1967.

Submarine clay slides are common in western Norwegian fjords. Early studies in the Hardangerfjord (Holtedahl 1965, 1975) unveiled multiple turbidites in the upper sediment column. Just 3 km south of Stryn a fairly resent slide was discovered in 1982 (Aarseth et al. 1989). Soundings by the Norwegian Hydrographic Office in May 1967 and April 1983 showed an increase in water depth of max. 55 m. Calculations of the two maps showed that sediment volumes of 15 x 10^6 m^3 had been removed during this time span.
A newspaper survey resulted in a report of a personal observation of three “inexplicable” waves up to 1 m high on the shore 10 km distal to the slide scar in September 1967. The telephone company also reported irregularities on a cable laid down in 1963. The slide scar is clearly seen on a resent TOPAS profile, Fig. 44 (Hjelstuen et al. in prep.).

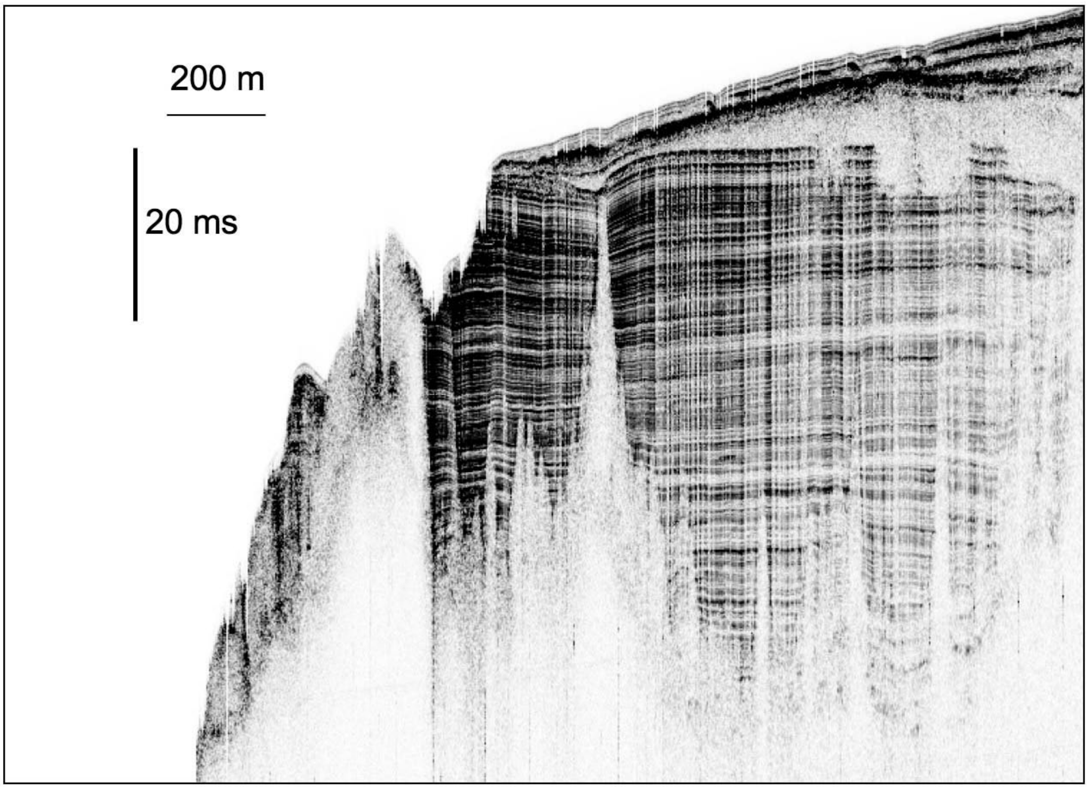

Fig. 44.TOPAS high-resolution seismic profile from inner part of Nordfjord, showing well-laminated glacimarine sediments and acoutic transparent slide debrites above bedrock. The 1967 Nordfjord Slide have created the 15 metres high slide scar observed within the glacimarine deposits.

Day 6: Tuesday August 19th Stryn - Erdalen - Stryn

Introduction

The last day of the excursion will bring us to two field areas in addition to the Jostedalsbreen National Park Centre. In Erdalen NGU (Landscape and Climate group) in cooperation with NTNU (Department of Geography) is running a long-term monitoring programme. The focus of this programme is on studying present-day rates and controlling factors of geomorphic processes, including both slope and fluvial systems.

Stop 1: Lower Erdalen valley (61° 51.69‘ N, 7° 08.38‘ E)

The U-shaped Erdalen valley is located close to Stryn in the innermost part of the Nordfjord area. The climate of this area is oceanic, in the upper parts of Erdalen subarctic oceanic, with a mean annual precipitation of ca. 1100 mm.
The valleys in the Stryn area are all heavily affected by glacier erosion. Erdalen is as a typical valley of the area deeply incised into bedrock, and the summits along the valley are up to 1200 – 1500 metres above the valley floor. The width of the valley floor varies along the valley but rarely exceeds 700 m. Bedrock is exposed along the valley bottom at several locations forming natural thresholds or sills controlling the incision of the river and creeks. In addition, local moraine ridges across the valley contain narrow pathways for fluvial channels. Bedrock is exposed along the valley sides and is alternating with numerous alluvial and colluvial fans, which partly form laterally connected aprons. The bedrock sills across the valley floor form boundaries of numerous interconnected valley basins. A small glacier lake is situated in the innermost valley basin at the margin of the present Erdalsbreen glacier, which forms an outlet glacier of the Jostedalsbreen ice cap.

Stop 2: Upper Erdalen valley (61° 51.22‘ N, 7° 09.32‘ E)

The upper Erdalen system can be subdivided into three areas: the braided sandur system, (Fig. 45) and the two tributary systems Vesledalen and Stordalen. Significant coupling between slope and fluvial systems exists today only in Vesledalen. The sub-recent and present-day net sediment budget of the braided sandur systems appears to be slightly negative. Erdalen terminates at lake Strynevatnet.
Present-day denudative processes include rockfalls, boulder falls, avalanches, slush flows, debris flows, creep, wash denudation, chemical denudation, glacial erosion and fluvial sediment transport. Postglacial modification of the glacial relief is altogether small which is due to the high resistance of the Precambrian gneisses in the area. The current slope processes generally cause a valley widening.
The runoff regime is complex, with runoff peaks occurring during snowmelt in spring, glacier melt in summer and heavy rainfall events. Fluvial bedload transport occurs to a significant extent only during high runoff events, and fluvial sediment transport in Erdalen is altogether supply-limited. The upper valley system is part of the Jostedalsbreen National Park and human impact exists only in form of grazing. Compared to that, farming and agricultural land use are dominant in the lower section of the valley.

The Erdalen Event

Following the Younger Dryas and early Preboreal, the first evidence of glacial activity in southern Norway in response to climatic change occurred around 10,000 cal. yr BP, when terminal moraines from the Erdalen Event formed up to 1 km beyond the position of later ‘Little Ice Age’ moraines.

Evidence for the timing of this early-Holocene glacier episode was first obtained in Erdalen, where the glacier re-advance left two sets of marginal moraines at Vesledalssetra, termed the Erdalen Event by Nesje et al. (1991). A basal radiocarbon date in a peat bog 300 m proximal to the outer terminal moraine yielded an age of 8810 ± 130 ^{14}C yr BP or 9875 (10,155-9560) cal. yr BP.

In front of present outlet glaciers on both sides of the Jostedalsbreen ice cap, sets of pre-‘Little Ice Age’ marginal moraines have been morphostratigraphically correlated with the

Erdalen Event. Radiocarbon dates from a peat bog in front of Nigardsbreen indicates that the first Erdalen Event re-advance culminated approximately 10,100 cal. yr BP, whereas the second occurred close to 9700 cal. yr BP (Dahl *et al.*, 2002).

Fig. 45. The sandur in Erdalen at Stop 3, Day 6. Photo: A. Beylich.

Stop No 3: Jostedalsbreen National Park Centre (61º 54.69' N, 7º 2.91' E)

Jostedalsbreen National Park Centre is situated beside the lake Strynsvatn. It was opened in 1993 and is owned by a private foundation based on idealistic objectives. The main building at the Centre has been constructed in a manner similar to Viking longhouses where pillars rather than the walls are supporting the roof. The longhouse has about the size of the biggest longhouse found in Norway. As a contrast to the old longhouse is the Cinema built in modern style, covered by Norwegian polished stone, Larvikitt. This symbolizes the meeting between the past and the future. In the cinema is shown a panoramafilm from Jostedalsbreen glacier. Animations from maximum of the last iceage that shows the landscape in Stryn 20.000 years ago and how the glacier completely melted away and was born again in a cold period ca. 6000 years ago. The film also tells how local people have used the glacier in the past and nowadays. Jostedalsbreen National Park Centre has been authorized by the Department of Environment to serve as the official visitor centre for the national park. The exhibitions contain both cultural and natural history. The latter includes animal life, avalanches and geology. Outrside there is a botanical garden as well as a geology park with bedrocks from all the 19 counties of Norway as well as the communes in the Sogn and Fjordane County.

Day 7: Wednesday August 20th Stryn – Oslo Airport – Oslo

This day we will start 08:30 to reach Oslo airport and Oslo in time for some of the participants to reach transport back home. There will be a lunch stop on the way (packed lunches from the hotel). Anticipated arrival Oslo airport Gardermoen at 16:00 and Oslo Central Station before 17:00.

Acknowledgements

Brian Robins has corrected parts of the text and Eva Bjørseth has assisted in making the final version of the figures.

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