

# Late Weichselian and Holocene glacier fluctuations along a south-north coastal transect in Norway

Climatic and methodological implications



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2004



**ISBN 82-497-0207-7**  
Bergen, Norway 2004

Printed by Allkopi, Bergen – ph.+ 47 55 54 49 40



## **Preface**

This Doctor Scientiarum thesis is a result of **NOR**wegian **Past Environments & Climate** (NORPEC), a strategic university program at the University of Bergen coordinated by Professor H. John B. Birks. As a Dr. scient student, I have taught under-graduate and master-students 25% of my time (1 year) paid by the Faculty of Social Science at the University of Bergen. I have spent my study time both at the Department of Geography and at Bjerknes Centre for Climate Research at the University of Bergen. I am grateful to Associate Professor Dr. Svein Olaf Dahl and Professor Atle Nesje who supervised this project. This thesis would not have been fulfilled without their encouraging participation. I am looking forward to continue our common struggle for a better understanding of past and future changes of the climate system.

Several other persons have been of great help for the long-lasting process of finishing my thesis. I am especially grateful to Joachim Riis Simonsen, Åsmund Bakke, Morten Diesen, Bjørn Kvisvik, Lars-Ivar Folgerød, Roy Sjonfjell and Sigurd Sandvold.

I would also like to express my sincere gratitude to all colleges at NORPEC for many stimulating meetings and discussions. Professor John Birks has been an excellent coordinator of NORPEC, chaired our discussions and kept an eye on the finances. Thanks to Anne Bjune for sharing the secrets of pollen and plants, and to Professor Reidar Løvlie for introducing me and the other NORPEC members to the mysteries of environmental magnetism.

To the technical and administrative staffs at the Department of Geography, Kjell Helge Sjøstrøm, Kjell Totland, Gro Aase, Mette Handler, Dagny Medaas Stiansen and Lillian Eide, and at the Bjerknes Centre for Climate Research, Rune Grimstad, Charla Melander and Gunn Holko, I give my warmest acknowledgement for always being helpful.

My colleagues and friends Øyvind Paasche and Øyvind Lie deserve a special thank for enthusiastic discussions, litres of coffee, late nights and for never giving up the idea of exploring and falsifying the nature of science.

I also want to thank my closest friends Bård, Ivar, Margunn and my brother Åsmund. Your patience with me, always busy, is beyond apprehension.

But most of all, thanks to Kristina. Without your everyday inspiration and enthusiasm during numerous days in the field, this thesis would maybe never have been finished.

Jostein Bakke, 18<sup>th</sup> February 2004.







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# **Part I**

## **Introduction**

**Late Weichselian and Holocene  
glacier fluctuations along a  
south-north coastal transect in  
Norway**

**Climatic and methodological implications**





# Late Weichselian and Holocene glacier fluctuations along a south-north coastal transect in Norway

## Climatic and methodological implications

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## Practical introduction

This thesis is divided into two parts. Part one is a general introduction to climate and methods, and part two is a presentation of the papers. The Introduction is divided into five parts. First, there is a general overview over different sources for palaeoclimatic information. Secondly, the main objectives are outlined. Then the climate in the North Atlantic region and its relation to glaciers at the west coast of Norway is discussed. Further there is a general introduction to reconstruction of glaciers and quantification of winter precipitation. As this thesis outlines both methodological and climatic issues, this is discussed in more general terms during the Introduction with respect to some of the results presented in the thesis. Part II is a presentation of four papers submitted to international peer-review journals. Paper 1 reviews conceptual methodological approaches for the reconstruction of glacier variations. Paper 2 presents some climatic implications of a Lateglacial and early-Holocene glacier reconstruction from Folgefonna. Paper 3 suggests some new methodological approaches for high-resolution glacier reconstruction in high alpine areas. Finally, Paper 4 outlines Lateglacial and Holocene glacier fluctuations and palaeoclimatic variations in Lyngen, northern Norway.

## Papers in Part II:

- Dahl, S. O., Bakke, J., Lie, Ø. and Nesje, A. 2003. Reconstruction of former glacier equilibrium-line altitudes based on proglacial sites: an evaluation of approaches and selection of sites. *Quaternary Science Reviews* 22, 275-287.
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- Bakke, J., Lie, Ø., Nesje, A., Dahl, S. O. and Paasche, Ø. (submitted). A high-resolution Holocene glacier reconstruction based on physical sediment parameters from proglacial lakes at northern Folgefonna, western Norway. *The Holocene*.
- Bakke, J., Dahl, S. O., Paasche, Ø., Løvlie, R. and Nesje, A. (Submitted). Late Weichselian and Holocene glacier fluctuations, equilibrium-line altitudes and palaeoclimate in Lyngen, northern Norway: a high-resolution composite chronology based on lacustrine and morpho-stratigraphical evidence. *Quaternary Science Reviews*.

## Introduction

The Holocene epoch followed the last major pulse of glaciation (the Younger Dryas period) at the end of the last ice age (Weichselian), and encompasses a period of time before there was any substantial anthropogenic climate forcing. The Holocene has up to quite recently been regarded as a period of relatively stable climate. Recently published evidence suggests, however, that the Holocene climate has been more variable than previously thought. The early to mid-Holocene was characterised by relatively warm conditions in the northern Hemisphere, in many areas warmer than in the 20<sup>th</sup> century (e.g. Alverson *et al.*, 2003). In the later part of the Holocene, temperatures have generally declined. This decline was punctuated by centennial-scale temperature oscillations, with the most recent cold episode (~AD 1550-1920), including the 'Little Ice Age', being one of the coldest periods in the entire Holocene in most regions.

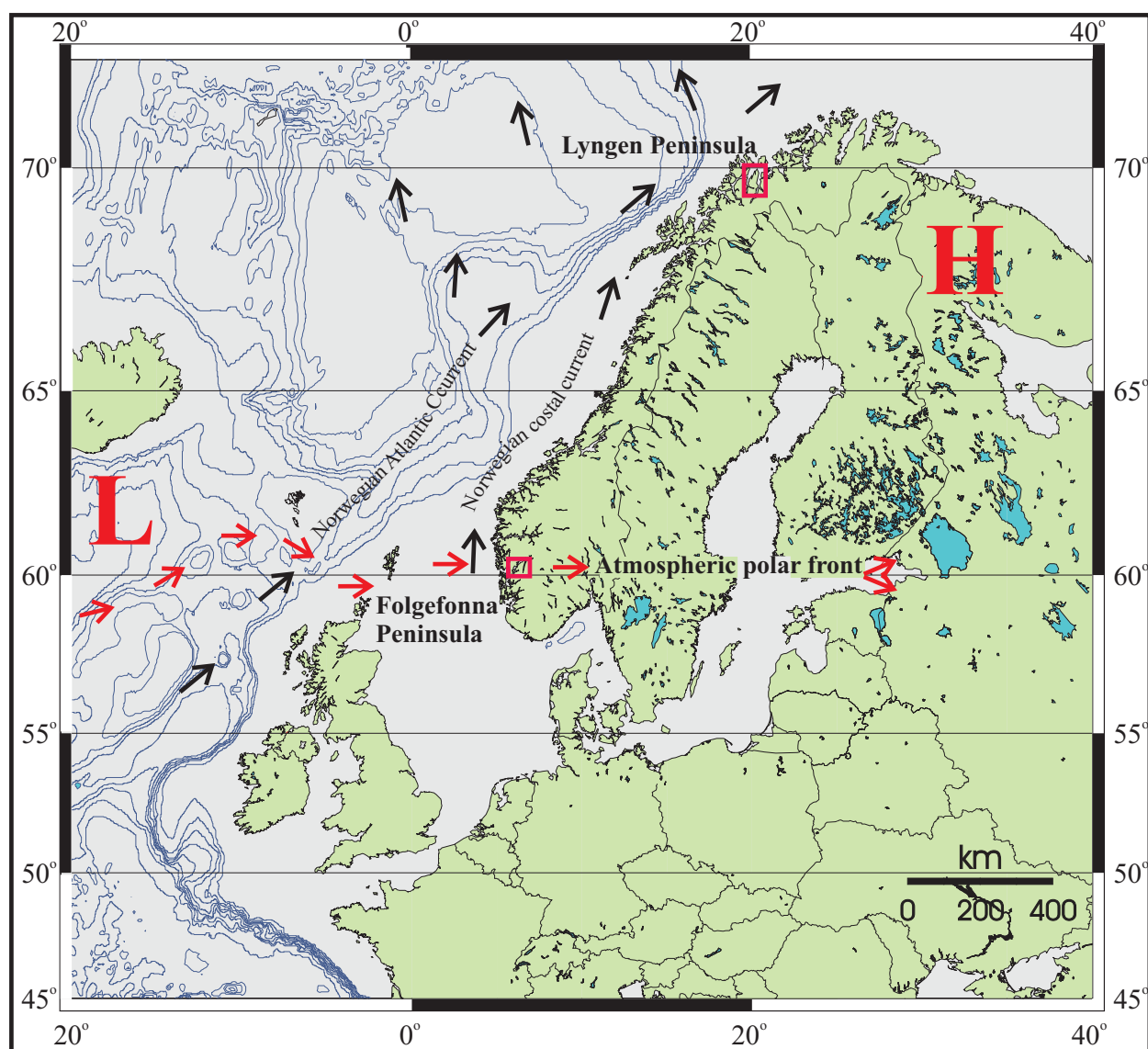
Holocene climate variations in Europe and the North Atlantic region (Figure 1) have been reconstructed from ice cores (Dansgaard *et al.*, 1993; Grootes *et al.*, 1993; O'Brian *et al.*, 1995; Stuiver *et al.*, 1995; Alley *et al.*, 1997; Dahl-Jensen *et al.*, 1998; Johnsen *et al.*, 2001; Fisher and Koerner, 2003), alpine tree-limit fluctuations (Kullman, 1981, 1995; Eronen *et al.*, 1993; Berglund *et al.*, 1996; Karlén and Kuylenstierna, 1996) glacier variations (e.g. Karlén, 1976, 1988; Nesje *et al.*, 1991; Dahl and Nesje, 1992, 1994, 1996; Karlén and Matthews, 1992; Matthews and Karlén, 1992; Leeman and Niessen, 1994; Nesje *et al.*, 1994, 1995; 2000a, 2001; Karlén *et al.*, 1995; Snowball and Sandgren, 1996; Svendsen and Mangerud, 1997; Matthews *et al.*, 2000; Snyder *et al.*, 2000; Barnekow and Sandgren, 2001; Dahl *et al.*, 2002; Seierstad *et al.*, 2002; Winkler *et al.*, 2003), pollen, chironomids and diatoms (Barber *et al.*, 1999a; Barnekow, 2000; Rosén *et al.*, 2001; Seppä and Birks, 2001; Bigler *et al.*, 2002; Korhola *et al.*, 2002; Seppä *et al.*, 2002a; Seppä and Birks, 2002; Seppä *et al.*, 2002b; Bigler and Hall, 2003; Davis *et al.*, 2003; Heikkilä and Seppä, 2003; Hannon *et al.*, 2003; Heiri *et al.*, 2003), plant macrofossils (Birks, 1991; Birks and Ammann, 2000; Birks *et al.*, 2000; Hannon *et al.*, 2003), tree rings (Briffa *et al.*, 1992; Schweingruber and Briffa, 1996; Briffa *et al.*, 1998; Kalela-Brundin, 1999; Kirchhefer, 2001; Briffa *et al.*, 2002; Grudd *et al.*, 2002; Helama *et al.*, 2002; Thun, 2002), oxygen-isotope records from lake sediments (Grafenstein *et al.*, 1998; Shemesh *et al.*, 2001; Hammarlund *et al.*, 2002; 2003), physical sediment parameters, such as loss-on-ignition and mineral magnetic properties, in lake sediments (Snowball *et al.*, 1999; Willemse and Törnquist, 1999; Battarbee *et al.*, 2001; Nesje and Dahl, 2001; Battarbee *et al.*, 2002; Kaplan *et al.*, 2002; Lotter and Birks, 2003; Rubensdotter and Rosqvist, 2003), peatlands (e.g. Chambers and Charman, 2004), speleothems (Lauritzen, 1996; Lauritzen and Lundberg, 1999),

marine proxies (e.g. stable isotopes, faunal variations, physical sediment parameters) (Lehman *et al.*, 1991; Koc *et al.*, 1993; Hafliðason *et al.*, 1995; Bond *et al.*, 1997, 2001; Fronval and Jansen, 1997; Klitgaard-Kristensen *et al.*, 1998; Barber *et al.*, 1999b; Bianchi and McCave, 1999; Grøsfjeld *et al.*, 1999; Hald *et al.*, 2001; Klitgaard-Kristensen *et al.*, 2001; Mikalsen *et al.*, 2001; Birks and Koc, 2002; Husum and Hald, 2002; Andersson *et al.*, 2003; Andrews and Giraudeau, 2003; Oppo *et al.*, 2003; Risebrobakken *et al.*, 2003; Sarnthein *et al.*, 2003; Solignac *et al.*, in press), marine molluscs (Salvigsen *et al.*, 1992; Hjort *et al.*, 1995; Salvigsen, 2002), and historical/ instrumental evidence (e.g. Pfister *et al.*, 1999; O'Sullivan *et al.*, 2002; Nordli *et al.*, 2003)

Continental climate records from Scandinavia capture Holocene insolation changes as well as changes in the North Atlantic Ocean and the atmospheric circulation. The proximity of this region to Greenland allows the comparison between Greenland ice-core records and continental records of northern Europe. In order to make reliable and likely predictions of future climate, and to separate natural from human-induced climate variability, it is important to know the rate and magnitude of past climate changes (e.g. Alverson *et al.*, 2003). Because instrumental meteorological records commonly are too short to cover the entire climate variability, the climate history for earlier periods has to be reconstructed from indirect (proxy) indicators. Previously, climate reconstructions from biological sedimentary remains were based on single indicator species or assemblages of taxa, and the results were usually interpreted in a qualitative and descriptive manner. Numerical techniques and approaches are now available that allow quantitative reconstructions from floral and faunal assemblages (ter Braak and Juggins, 1993; Birks, 1995, 1998, 2003).

A major obstacle for producing reliable predictions of global climate change and its environmental impacts is a lack of data on time scales longer than the instrumental records. Natural archives of past climate variability can provide relevant information over longer timescales. Furthermore, a better understanding of past variability can be used to improve the performance of climate models. Model experiments can then again be used to investigate the causes and dynamics of past climate variability.

The latest research on recent climate variability is increasingly forced towards the view that greenhouse gas forcing is becoming the dominant, though not the only process driving the warming trend over the last 60 years or so. At the same time, detailed analyses of ice core records from glacial terminations show how closely atmospheric greenhouse gas concentrations track rapid warming, probably by acting as a major feedback mechanism during the transition to interglacial conditions. These and other lines of evidence from studies of the



**Figure 1** Location map of the North Atlantic region. Study areas are shown in red frames. Red arrows indicate the average position of the atmospheric polar front. Black arrows indicate the main ocean currents of the modern North Atlantic region. The low-pressure field south of Iceland and the high-pressure field over eastern Scandinavia are of importance for the climatic development at the west coast of Norway.

present-day atmosphere reinforce the view that future climate change will be driven by complex interactions between the effects of anthropogenically produced greenhouse gases and the effects of natural climate forcing factors (e.g. solar variability and volcanic aerosols). It is therefore an urgent need to improve our documentation and understanding of natural climate variability for periods stretching back beyond the instrumental records.

Natural climate variability, whether associated with mechanisms external or internal to the Earth climate system, is expressed on inter-annual, decadal, century, and millennial time-scales. Climate variability on these time scales and its interactions with future anthropogenic forcing are of great importance to human society. The need for a strong focus on high-resolution records of climate variability on these time scales is highlighted by the growing extent to which climate and earth system modellers are beginning to run experiments that seek to

replicate variability on these time scales.

The urgent need to identify and understand the nature of extreme weather events and inter-annual to century-scale climate variability, reflects the need to inform and to evaluate the developing generation of coupled ocean-atmosphere climate models. Knowledge of past climate variability on these time scales needs to be gained both from long-term instrumental records and from well-calibrated proxy data derived from natural archives.

In ideal circumstances such archives provide accurate records of climate history, and can be dated with annual to decadal precision on a calendar year time scale. Ideally such records can also be inter-correlated through time and extended to the past to include the full range of climate variability of relevance to climate predictions. Commonly, however, no single climate archive possesses such properties, and climate reconstructions consequently need to combine information from different sources





**Figure 2** 3D-illustration showing the Lyngen Peninsula in northern Norway. The red dot indicates lake Aspvatn described in paper 4. (Illustration: Kjell Helge Sjøstrøm).

within and between geographical regions.

Resolving these methodological issues will allow us to tackle some of the more important questions relating to climate variability on these time scales that have been identified in international programmes like IGBP, PAGES and HOLIVAR.

This thesis sustain the above discussed proxy records from the Late Weichselian and Holocene by reconstructing glacier variations and changes in winter precipitation at the west coast of Norway.

## Objectives

The main objectives of this thesis can be listed as followed:

### Palaeoclimatic objectives

- to evaluate glacier variations along a south-north transect at the west coast of Norway during the Lateglacial and Holocene (Figures 2 and 3)
- to transform glacier variations into former winter precipitation and thereby provide a reliable proxy for the winter climate in the North Atlantic region
- to analytically examine possible large scale atmospheric circulation patterns backwards in time beyond the historical records
- to evaluate early-Holocene glacial events in order to identify possible mechanisms for abrupt

climate change during the time span

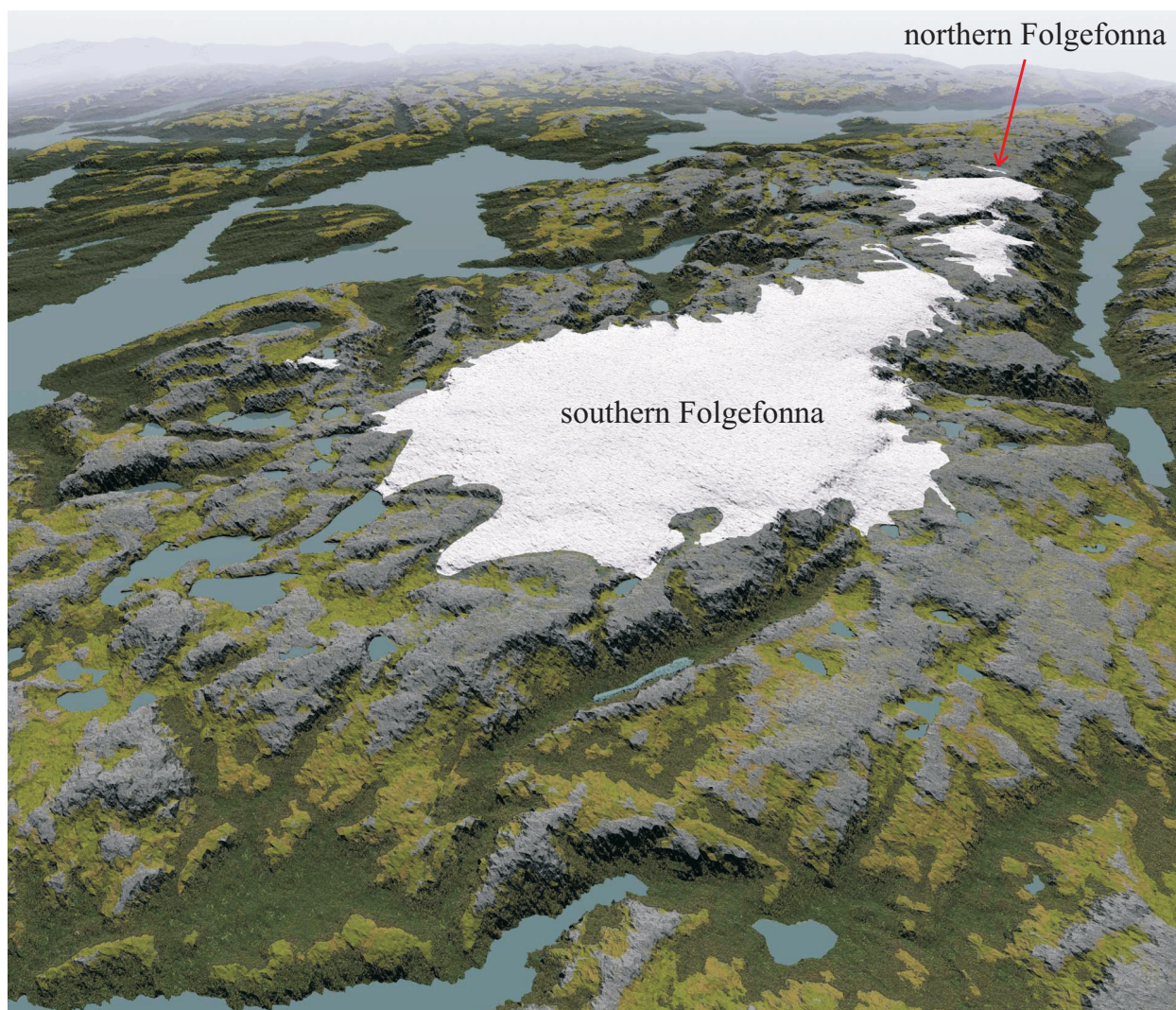
### Methodologically objectives

- to develop a methodological approach to be applied on lake sediments in areas with low organic production
- to use physical sediment parameters as a direct measure of glacially derived sediments deposited in proglacial lakes for quantification of former glacier size
- to provide data sets of decadal to multi-decadal time scale of former glacier variability

### Climate development at the west cost of Norway – glaciers as a palaeoclimatic indicator

This section deals with the coupling between glacier fluctuations and climate at the west coast of Norway. It is therefore necessary to describe the overall forcing mechanisms for the climate in the North Atlantic region at present. This is done by a brief introduction to the atmospheric and oceanographic components influencing the climate along the west coast of Norway. This section is further dealing with the transformation from a change in glacier size towards reconstruction of winter precipitation.





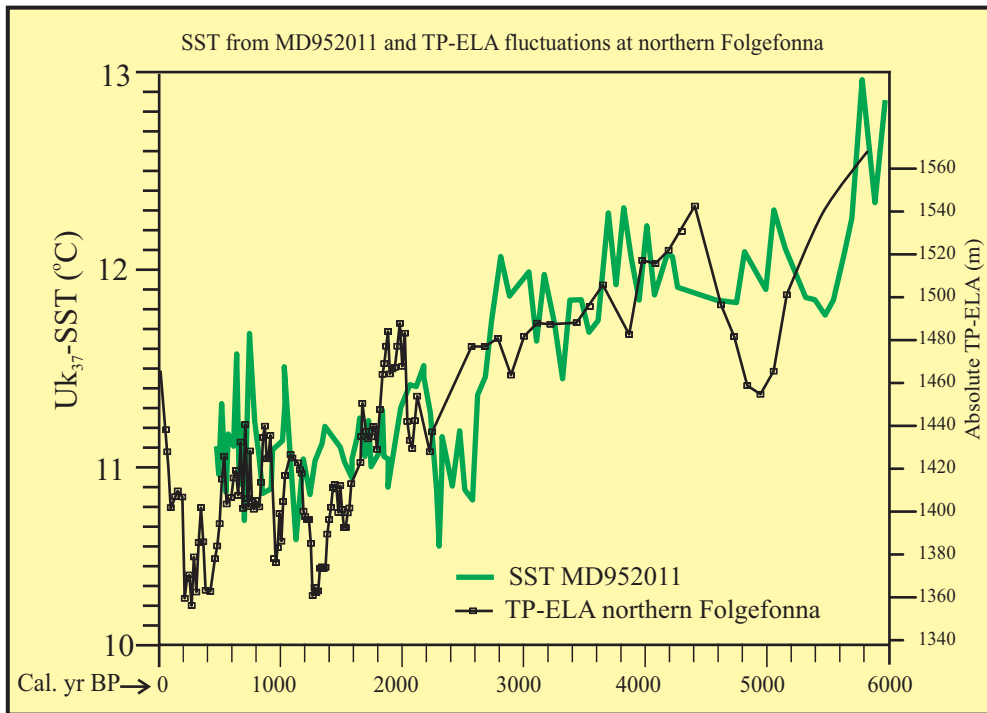
**Figure 3** 3D-illustration showing the Folgefonna Peninsula in south-western Norway. Study area in paper 2 and 3 is at the north side of the ice-cap northern Folgefonna (red arrow) (Illustration: Kjell Helge Sjøstrøm).

### Climate at the west coast of Norway

As the climate system is a dynamic interaction between the oceans and the atmosphere, and their feedback mechanisms, it is a system of great complexity. In the northern hemisphere a major element is the ocean heat advection towards higher latitudes, which, for this region, has a large impact on adjacent landmasses. Warm Atlantic water flows across the northern North Atlantic, around Iceland and along the eastern boundary of the Nordic seas. This circulation contributes with heat and moisture to north western Europe and Iceland. On its way, heat is gradually released, and the density of surface water increases and sinks to form North Atlantic Deep Water (NADW), a driver of the global thermohaline circulation (Broecker, 1991). Changes in the ocean circulation of the North Atlantic, and convective renewal of NADW could thereby also have global consequences. This highlights the importance of the North Atlantic region as a study area for climate change. The nonlinear ocean circulation system has alternated between times of strong thermohaline circulation of the North Atlantic and a total shutdown of shallow ventilation (Mababe

and Stouffer, 1988; Broecker, 1991). Freshwater pulses during glacial times are considered as possible triggers causing perturbations to the present day ocean circulation.

Another important contributor to the climate variability in the terrestrial systems at higher latitude is the atmospheric pressure patterns (Hurrell, 1995). This is highly influenced by the atmospheric polar front appearing at 60 degrees north, where cold dry air from the arctic region is mixed with warm humid air from lower latitude. Caused by this mixing the circulation at the west coast of Norway is characterised by migratory cyclones within the North Atlantic zone of westerlies, causing the prevailing westerly and south-westerly winds. These air masses release their humidity as they enter the rugged mountains of western Norway giving orographically enhanced frontal precipitation. As a consequence, the mean annual west-east precipitation gradient in southern Norway is ~2000 mm, comparing Bergen at the west coast with Folldal in the central inner parts of eastern Norway. The strength of the westerlies is strongly linked to the North Atlantic Oscillation index (NAO), which quantifies the pressure difference between the subpolar low-pressure system



**Figure 4** SST record from core MD95-2011 (Calvo *et al.*, 2002) together with the TP-ELA reconstruction at northern Folgefonna. The two-stepped cooling in SST during the late-Holocene is recognized more-or-less simultaneously as a two stepped lowering in the TP-ELA at northern Folgefonna (Paper 3).

near Iceland and a subtropical high-pressure system near the Azores (Hurrell, 1995; Hurrell *et al.*, 2003). The strong association with the 'Icelandic low' is well indicated by the robust correlation of  $r = 0.9$  (based on a correlation of data between 1820 and 2000). The index describes the dominant mode of winter climate variability in the North Atlantic region, ranging from central North America to Europe and much into northern Asia. Of importance for the weather at the west coast of Norway is a high pressure field lying over Scandinavia (Gulf of Bothnia) or western Russia. This anticyclonic pattern may block the prevailing westerly wind-field. In this situation a quasimeridional circulation replaces the zonal circulation of the North Atlantic, and the weather in southern Norway is controlled by the position of the blocking anticyclone. This means that the westerly humid air can be directed either to the south or to the north of southwestern Norway, giving increased precipitation in eastern or northern Norway.

The mechanism coupling the atmosphere and the oceans are not well understood. However, an ensemble of atmospheric general circulation model (AGCM) simulations demonstrate that winter climate is linked to the temporal history of SST in the North Atlantic region (Feddersen, 2003; Hurrell *et al.*, 2003). This may indicate that the atmosphere is closely related to changes in the oceans and even force some of the long term trends in NAO variability.

Several events during the Holocene are recognized in proxies both from the oceans and from proxies in the terrestrial system. An example is the widespread 8.2 event (termed 'Finse Event' in southern Norway), which is seen as a glacial event, loss-on ignition reduction in lake sediments,

temperature lowering in the ocean and in the ice cores from Greenland (e.g. Dansgaard *et al.*, 1993; Dahl and Nesje, 1996; Klitgaard-Kristensen *et al.*, 1998; Nesje and Dahl, 2001). Release of fresh water from the down wasting Laurentide ice sheet is suggested as a possible explanation. The link between the oceans and the atmosphere is also seen during the last 6000 cal. yr BP where e.g. sea surface temperatures (SST) from the Vøring Plateau (Calvo, *et al.*, 2002) and glacier growth in southwestern Norway show some of the same trends (Figure 3). It can be assumed that the boundary conditions for glacier growth and SST are forced by the same mechanism leading to a change in the atmospheric and oceanic conditions. The explanation for this coupling is not shown in other mechanisms than in orbital forcing. The changes in boundary conditions might therefore be linked to changed insolation over the northern Hemisphere (Bond *et al.*, 2001).

The changes in atmospheric circulation and changes in the ocean circulation are reflected in glacier variations in western Scandinavia. As the mass balance of a glacier mainly is the result of winter accumulation, summer temperature and prevailing wind direction it can reflect changes in wind directions, summer and winter climate (Dahl and Nesje, 1992). It is possible to use glaciers in different climate zones and with different aspect to reflect either one of the above climatic factors affecting the glacier.

#### **Reconstruction of the equilibrium-line altitude (ELA) and palaeoclimate**

The key objective of this thesis was to reconstruct former glacier variations and hence variations in ELA

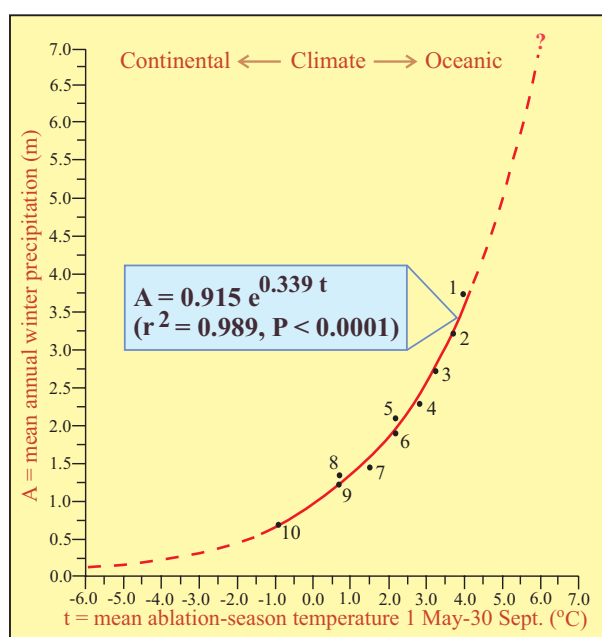


at two sites along the west coast of Norway (at the Folgefonna and Lyngen peninsulas). The definition of ELA is the altitude of a theoretical line that defines the altitude where annual accumulation equals the ablation. Changes in the ELA is regarded as the most useful parameter to quantify the influence of climate variability on glaciers (e.g. Andrews, 1975; Porter, 1975). Dahl and Nesje (1992) introduced the terms temperature-precipitation equilibrium-line altitude (TP-ELA) and temperature-precipitation-wind equilibrium-line altitude (TPW-ELA) to distinguish between glaciers which have low dependence on wind-accumulated snow (such as plateau glaciers) and glaciers which are dependent on the additional accumulation from wind-blown snow (such as cirque glaciers). The glacier ELA in Scandinavia is lowest in the west (c. 900-1200m), increasing towards the east until ELAs reach more than 1700 m and 2000 m in the Sarek mountains and in eastern Jotunheimen, respectively. Cumulative mass balance records during last 40 years have revealed diverging trends between glaciers in maritime and continental climate regimes of southern Norway, where the maritime glaciers have gained mass and the eastern, continental glaciers have lost mass during the last 40 years. At the west coast of Norway, the glaciers are mainly reflecting winter precipitation, whereas the glaciers in continental areas are mainly reflecting ablation-season temperature. The continental glaciers of southern Norway are similar to the high-latitude glaciers in having a primary summer balance/net balance (Bs/Bn) correlation, whereas the maritime glaciers at the west cost of Norway have a primary winter balance/net balance (Bw/Bn) correlation. It is also shown some linkages to the North Atlantic Oscillation in the Bn of maritime glaciers in western Norway (Nesje et al., 2000b).

By transforming former glacier fluctuations to winter precipitation (Dahl and Nesje, 1996), we can use glacier fluctuations as a helpful palaeoclimatic tool. Winter precipitation based on high-resolution glacier fluctuations is one of few reliable proxies for winter climate in the terrestrial system. Maritime glaciers where Bn is mainly controlled by the winter weather, e.g. at the west coast of Norway, are highly appropriate in this context.

Stable oxygen isotope records on bulk carbonates from Sweden may perform some of the same patterns in humidity as the winter precipitation records from western Norway (Hammarlund et al., 2003). It is, however, difficult to fully isolate the winter component using stable oxygen studies of carbonate-rich sediments.

From northern Norway there are some reconstructions of mean annual temperature (MAT) using speleothems from caves (Lauritzen and Lundberg, 1999). These data may be regarded as having a winter season component, but it is difficult to isolate the winter season from the whole year. From NW Scotland luminescent organic matter in stalagmites are used as a proxy for precipitation. This



**Figure 5** Close non-linear (exponential) relationship between mean ablation-season temperature (1 May – 30 September) and annual winter precipitation at the equilibrium line of 10 Norwegian glaciers existing in different climatic regimes (1. Älftobreen, 2. Engabreen, 3. Folgefonna, 4. Nigardsbreen, 5. Tunsbergdalsbreen, 6. Hardangerjøkulen, 7. Storbreen, 8. Austre Memurubreen, 9. Hellstugubreen, 10. Gråsubreen) (Dahl et al., 1997).

high resolution record performs evidence linking the continuously banded stalagmite to the NAO because of the location of the cave (Proctor et al., 2000), and may be regarded as a liable proxy for winter climate over the North Atlantic region.

Another terrestrial proxy giving records of winter climate in the North Atlantic region is high-resolution glaciochemical time-series developed from subannual sampling of an ice core from central Greenland calibrated with instrumental series of atmospheric sea-level pressure records (Meeker and Mayewski, 2002). This demonstrates that glaciochemical proxies have a potential to record the behaviour of past atmospheric circulation.

### Quantification of winter precipitation

Two new records of winter precipitation are presented in this thesis, one from southern Norway and one from northern Norway. Both are based on recent temperature reconstructions by botanists in the NORPEC project. The Holocene winter precipitation record from southern Norway is fully presented in Bjune *et al.* (submitted) and from northern Norway in Bjune *et al.* (in press). As this method of retrieving palaeoclimatic information by combining geological and biological proxies is rather new it may be appropriate to clarify the method used in this thesis.

The method goes back to Ahlmann (1924), who was the first to indicate a general relationship between accumulation (A) and summer temperature (t) at the ELA. Based on mean ablation-season temperature (1 May-30 September) and winter

precipitation (1 October-30 April) at the ELA of 10 Norwegian glaciers situated in oceanic to continental climate regimes (Figure 5), a close non-linear (exponential) relationship has been demonstrated (O. Liestøl in Sissons, (1979)). The close exponential relationship between the two parameters is expressed by the regression equation:

$$A = 0.915e^{0.339t} \quad (r^2 = 0.989, P < 0.0001) \quad (1),$$

where A is in meters water equivalent and t is in °C. This relationship implies that if independent data for either winter precipitation or ablation-season temperature at the ELA is known, the other factor can be calculated. It also implies that if the former ELA is known, it is possible to calculate how the other parameters has fluctuated (Dahl and Nesje, 1996).

ELA fluctuations with time mainly results from the interaction of mean ablation-season temperature and winter precipitation, which are related to equation 1. Whereas temperature is lowered by a linear environmental lapse rate of 0.6°C/100m with increasing altitude, precipitation increases exponentially with altitude. Adjusted for land uplift, all deviations from present values in the former ELA or in summer temperature are converted into equivalent temperature units. The present mean ablation-season temperature is converted into equivalent for this deviation (°C). Corresponding winter precipitation at the present ELA is calculated by substitution in equation 1. An independent proxy for summer temperature can be obtained from different sources, whereas estimates of former winter precipitation at the present ELA may be adjusted with altitude by c. 8%/100 m (Haakensen, 1989). By substituting this on present day temperature and precipitation pattern in southern Norway, using above-mentioned temperature and precipitation gradients, the glacier distribution in southern Norway is reproduced (Lie *et al.*, 2003).

A common problem when obtaining continuous records of former winter precipitation is that all glaciers studied in Scandinavia (until present) has been melted away for shorter or longer time span. During periods with no glacier, the reconstructed winter precipitation is expressed as maximum *potential* winter precipitation before a glacier reform at the site. For continuous reconstructions of winter precipitation in this study we have used the altitude of instantaneous glaciation (AIG) adjusted for land uplift during has been used in periods without glaciers. This is defined as the altitude of the highest lying mountain plateau suitable for glaciation in the study area. The climatic information derived from periods without a glacier present is only interesting as *maximum* winter precipitation values. However, they can still be used in further climatic analysis with this important assumption in mind.

Retrieving reliable estimates for former ELAs can to some extent be problematic. In periods with advancing glaciers it is possible to calculate directly

the ELA based on the Accumulation Area Ratio (AAR) (Andrews, 1975; Dahl and Nesje, 1992). We have performed the AAR reconstruction by using a vector GIS program on N-50 map datum. This approach gives very accurate calculation of the former glacier area. Because of the topographic setting of the reconstructed glaciers in Lyngen (paper 4) we used 'Little Ice Age' ratio (LR) to reconstruct the ELA variations (Dahl *et al.*, 2002). This is done since a lowering of the ELA will not be reflected in the AAR, despite an increase in ice volume and an advance on the flat valley bottom.

During time spans with retreating glaciers it is difficult to identify the rise in ELA causing the glacial retreat. Another problem is when converting the reconstructed ELAs to winter precipitation. As eq. 1 is calculated from glaciers in steady state, winter precipitation estimates could be erroneous when adapting this equation to glaciers in either advancing or retreating states.

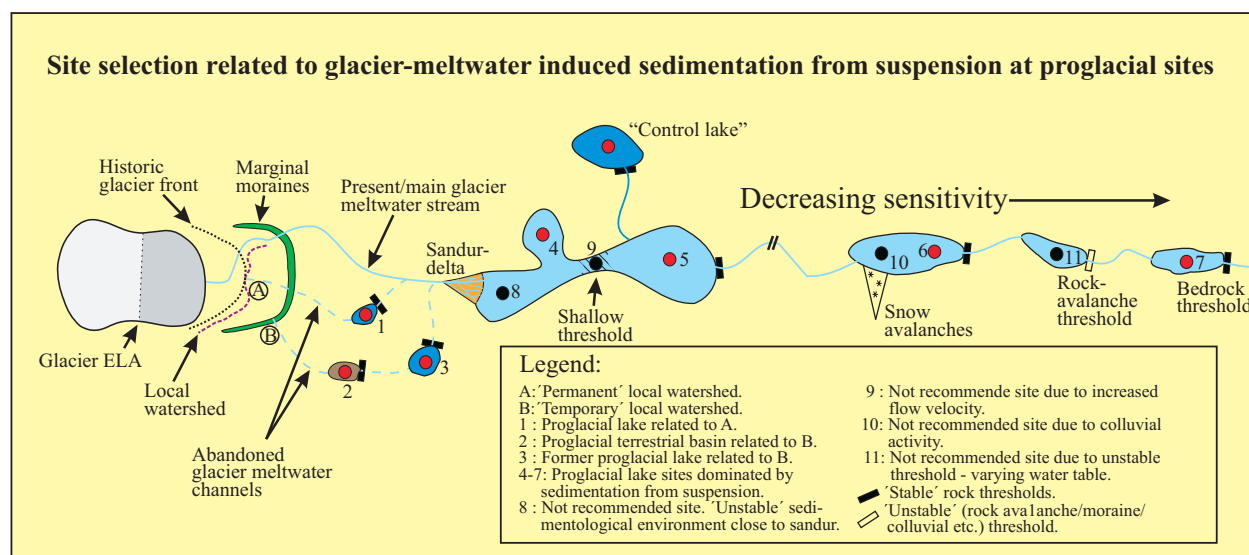
Temperature reconstructions based on pollen transfer functions are done with an error estimate of c. ±1°C (Birks, 2003). This affects all reconstructions of winter precipitation together with the above mentioned problems with ELA estimates. In paper 2 some of these possible error sources are taken into account when calculating winter precipitation. The result is rather large error bars in the reconstructed winter precipitation, especially when the ELA is rising after a defined glacier advance.

## Methodological implications inferred in this thesis

The results obtained in this thesis are based upon geomorphological mapping and studies of lake sediments. Two main issues are reviewed in this section; the importance of site selection in a catchment with glaciers (Figure 6) and analysis of physical sediment parameters such as bulk density and grains-size variations.

### The geomorphological setting and lake sediments

The main advantage of lake sediments in environmental research is the relatively easy way to adopt age control for catchment changes combined with the possibility for continuous glacier reconstructions. It is, however, several factors that potentially can complicate sedimentary records from lake sediments (Rubensdotter and Rosqvist, 2003). The first is to adopt control over possible catchment processes and processes within the lake. Site selection and careful mapping of the surrounding geomorphology are therefore of fundamental importance when working with sedimentary records. However, some studies seek to utilize catchment processes for reconstruction of geomorphology. It is nevertheless complicated as these may be related to climatic changes (such as temperature and precipitation) or processes of non-climatic relevance



**Figure 6** Potential sites for investigating variation in glacier activity/ELA in a schematic catchment with one glacier and a chain of proglacial lakes. Lacustrine and terrestrial sites related to both 'permanent' (bedrock) and 'temporal' (ice-marginal moraines) local watersheds, control lakes and various settings related to proglacial lakes are shown. Recommended coring sites are marked with red dots, whereas secondary sites close to unstable sedimentary environments (normally not recommended) are marked with black dots. See legend for further details (Paper 1).

influencing the lake sediments (such as rockfall, cattle grazing etc.). An initial site selection adapted to the phenomenon to be studied is therefore of great importance. A basic concept is to use simple systems that record a purified signal of the requested proxy. In this early phase of describing a research project it is of great importance to examine the topographic configuration and the geomorphological processes in a catchment.

For finding an area with proglacial sites suitable for reconstructing variations of former ELAs some basic factors should be in mind:

- Search for a catchment where a single glacier has existed throughout the Holocene.

- Search for a glacier where the component of wind driven snow accumulation is minimized. The ideal glacier is a small ice cap (plateau glacier) with outlet glaciers in all aspects. At this type of glaciers the influence of wind can be more or less be neglected.

- Search for temperate glaciers with proper sediment production at the base of the glacier, not polar or polythermal glaciers that are frozen to the bed.

- Search for an area with a number of downstream proglacial lakes and other sites suitable to document variations in the ELA.

- Search for proglacial lakes that are dammed by a rock sill and not by moraines, colluvial fans, rock avalanches etc.

- Search for lakes with a residence time long enough to allow suspended sediments to settle.

- Search for an area where there is representative marginal moraines of known age which can be used to calibrate the signal derived through the sediment parameters.

- Search for an area without superficial sediments and active geomorphological processes that may

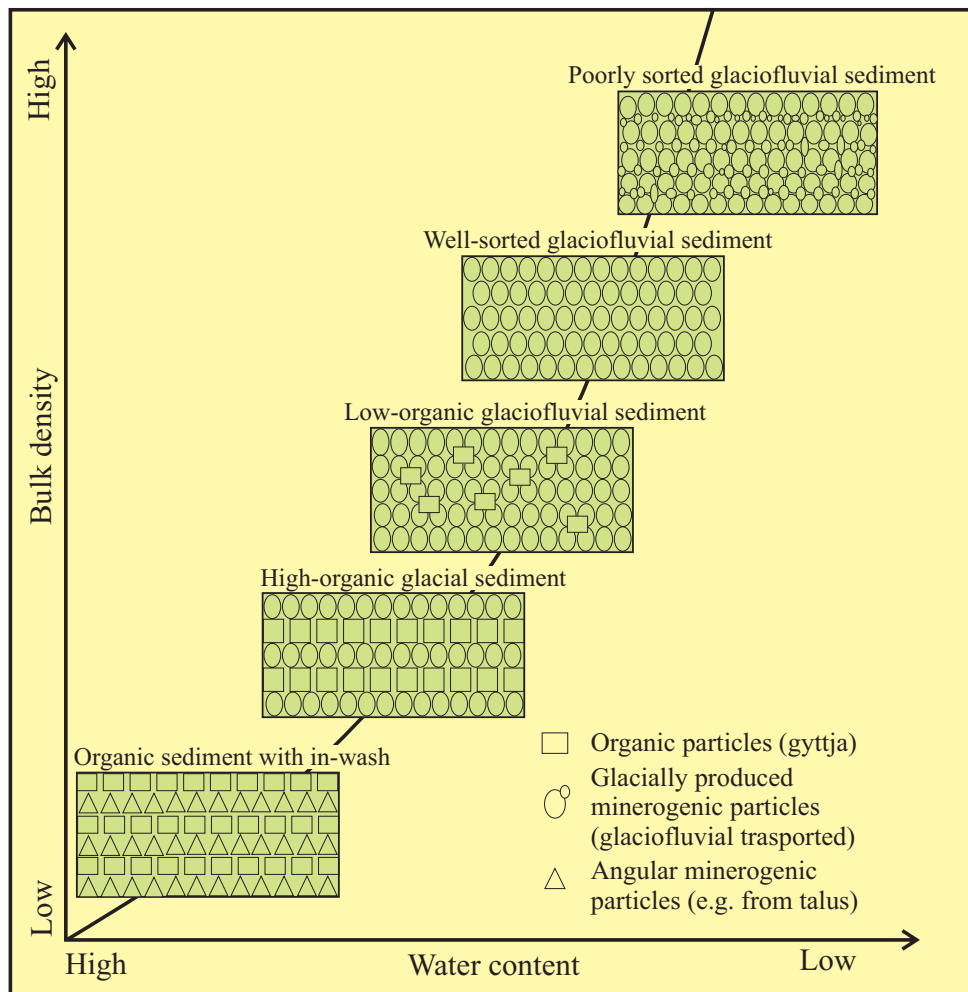
influence the lake sedimentation (e.g. snow avalanches, river floods, rock avalanching)

After retrieving the cores from a well-suited proglacial lake several laboratory analyses have to be done. Proper sampling strategies should be in mind from opening the core towards the final shipping of the radiocarbon dates to an external laboratory. Implications from this study are beyond measuring of loss-on-ignition (LOI), also to examine physical sediment parameters like wet and dry bulk density, water content, surface susceptibility and grain size analyses at high resolution. The advantages of remanent magnetisation and saturation isothermal remanent magnetisation are also successfully incorporated in the analyses based on grain size distribution and other physical parameters.

### Physical sediment parameters and its relation to former glacier size

Studies on modern Norwegian glaciers have shown that the sediment yield is positively correlated with glacier size (data in Roland and Haakensen, 1985). Measurements of the proportion of glacial material in proglacial lake sediments may therefore provide continuous records of glacier fluctuations. The use of lake sediments in this context is widely used in Scandinavia (e.g. Karlén, 1976, 1981; Leonard, 1985; Nesje *et al.*, 1991; Matthews and Karlén, 1992; Dahl and Nesje, 1994; Nesje *et al.*, 1995; Dahl and Nesje, 1996; Snowball and Sandgren, 1996; Matthews *et al.*, 2000; Nesje *et al.*, 2001). Various approaches related to proglacial sites are all using a conceptual model of glacier-meltwater induced sedimentation in which the minerogenic (non organic) component of the sediments is related to the presence of a glacier





**Figure 7** Schematic figure explaining the relationship between bulk density and water content related to type of sediment. Angular minerogenic particles give higher porosity than rounded glacier-meltwater derived minerogenic particles. Lowest bulk density values are obtained from sediments dominated by gyttja and angular minerogenic particles, whereas the highest bulk density values are obtained in poorly sorted glaciofluvial sediments (Paper 3).

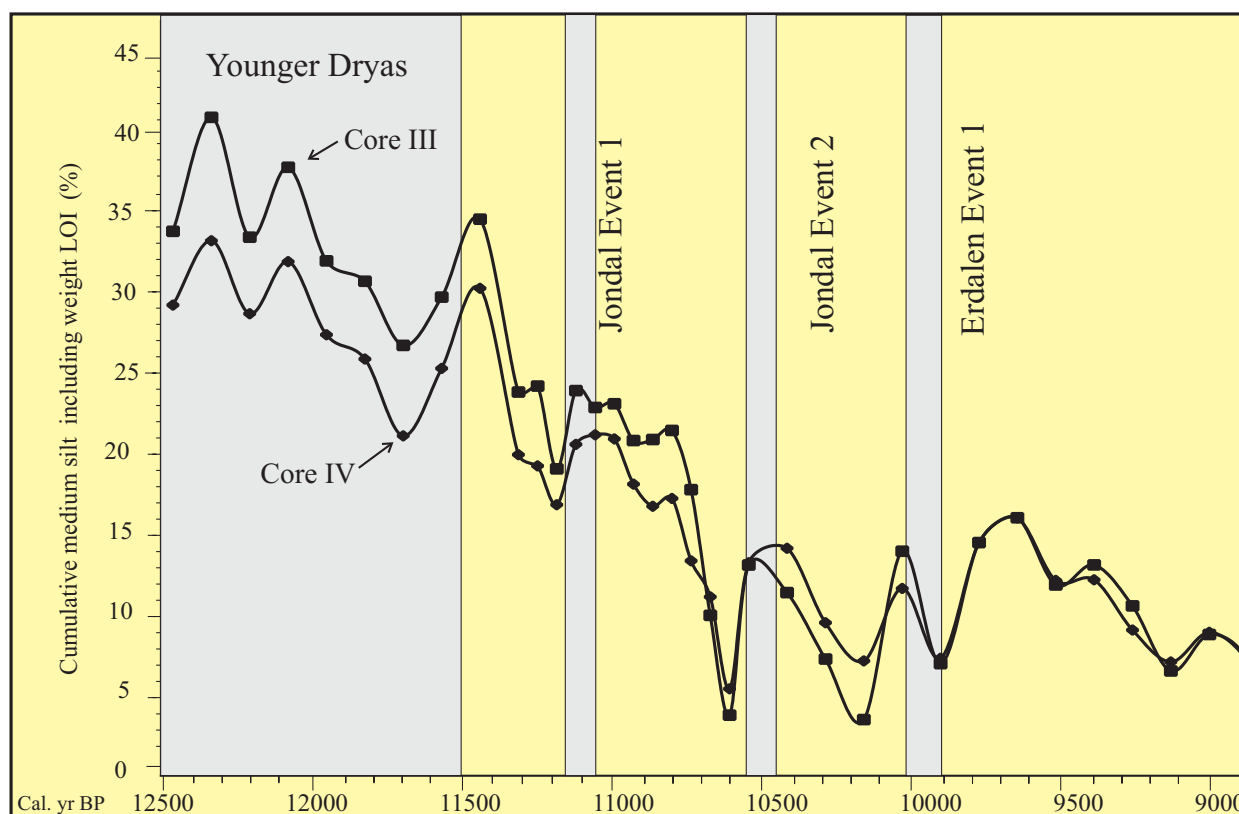
in the catchment (e.g., Karlén, 1981; Leonard, 1985; Dahl and Nesje, 1994; Nesje et al., 2001; Dahl et al., 2003).

However, only a few studies have examined the physical properties of the sediments in detail and especially the minerogenic material produced by the glacier (Leonard, 1985; Souch, 1994; Rosqvist, 1995; Snowball and Sandgren, 1996; Matthews *et al.*, 2000; Nesje *et al.*, 2000a; Nesje *et al.*, 2001; Lie *et al.*, In press). The most common approach is to use the organic matter (LOI) as an inverse indicator on inorganic deposition. In lakes with high minerogenic sedimentation and/or low organic production (~5%) this approach has its limitations since actual variations below this value will not be detected. We have, therefore developed an approach that quantifies glacier variations (and thus ELA variations) based directly on glacial sediments. Variations in physical sediment parameters in proglacial lakes are, however, mainly affected by production rates of sediments (controlled by glacier size) and transport in the hydrological system (runoff controlled by glacier size).

As the nature of glacial erosion is reflected by the

supply of insoluble particles to a river system, analyses of physical properties of the glacial sediments may be a diagnostic parameter for variations in glacier size. Warm-based glaciers produce abundant clay-silt size fractions that are transported downstream to produce characteristic signatures in glacio-lacustrine sediments (Østrem, 1975).

The use of grain-size variations have, however, not been widely used in this context. For calculating the absolute amount of minerogenic sedimentation in a lake, it can be useful to flux-correct the signal. A limitation for flux correction is commonly the number of obtained radiocarbon dates. Another important factor concerning the grain-size distribution in proglacial lakes is that glaciers do not normally produce one dominating particle size fraction. As seen from till studies, glaciers produce a composition of more-or-less all grain-sizes (Vorren, 1977). The glacial transport length and the size of the glacier do not seem to influence the grain-size distribution of glacial sediments (Jørgensen, 1977; Haldorsen, 1981, 1983). The grain-size variations in these sediments deposited in lakes are therefore



**Figure 8** Cumulative medium silt plotted versus loss-on-ignition for core III and IV in lake Vetlavatn (paper 2). The figure shows that the input of silt decreased after the Younger Dryas with small peaks during early-Holocene glacial events. Probably because of increased precipitation, the content of medium silt increased after the 'Erdalen Event 1' (Paper 2).

mainly reflecting changes in fluvial and lacustrine systems. As the transport and sedimentation in fluvial systems are closely related to Hjulströms diagram (Sundborg, 1956), high-energy streams deposit less fine grained sediments, and *vice versa*. In 'open-ended' lakes, the finest grain-sizes will be transported further downstream due to stronger currents and slow settling (this is shown from lake Aspvatn in paper 4). In a small, almost closed sediment basin, the grain-size distribution will consist of all grain-sizes suitable for suspension (1-63  $\mu\text{m}$ ), commonly giving more sediments per unit time than an 'open-ended' lake basins (shown in lake Dravladalsvatn, paper 3).

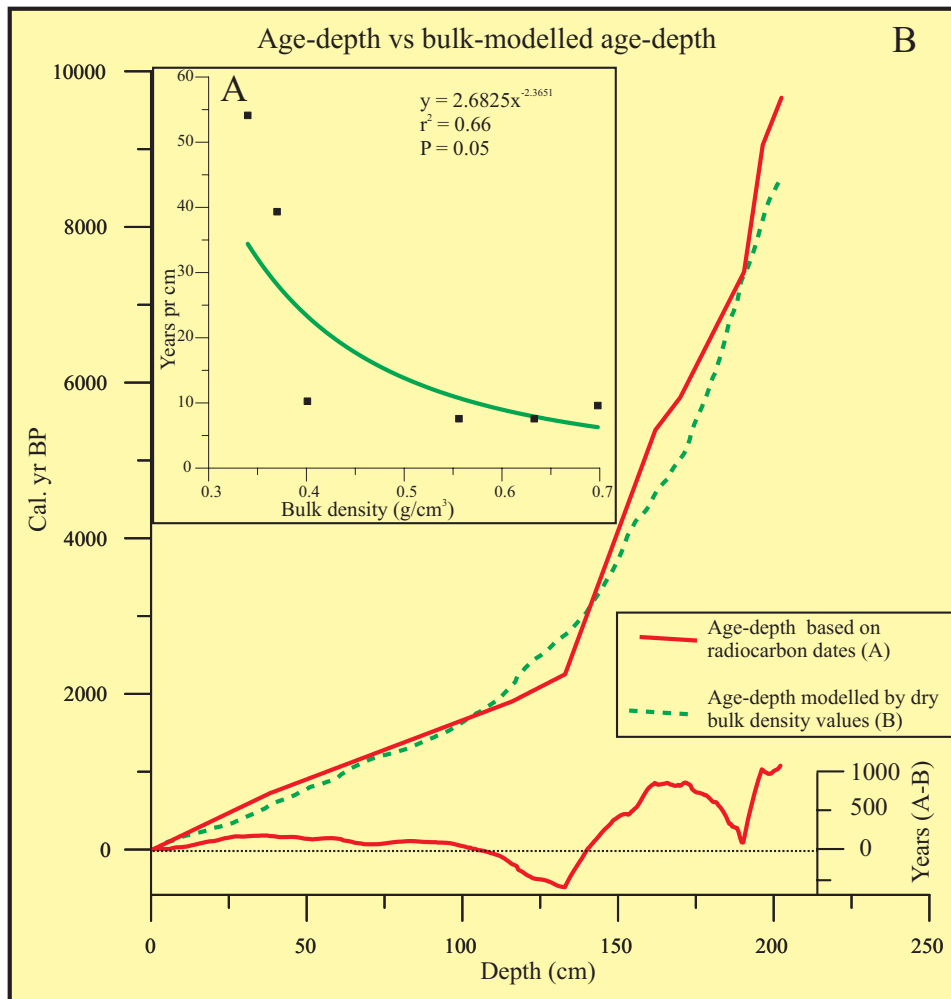
Bulk density acts as an additive parameter on the inorganic sedimentation. Bulk density express the ratio of the mass of dry solids to the bulk volume of a sediment (Blake and Hartge, 1986). Commonly, this parameter defines how granular, fibrous and powdery materials pack or consolidate under a variety of conditions and this parameter is therefore a reflection of porosity of the sediment (Figure 7). Changes in flux and packing (reflected in grain-size composition) are probably the most important parameter in a proglacial lake (Webb and Orr, 1997). Purely organic sediments should potentially be reflected by the lowest bulk values, whereas the highest values are expected in sediments consisting of poorly sorted minerogenic sediments. Most of the source material in a proglacial lake has the same

origin (glacially derived) and is therefore neglected as a variable factor for the bulk density values. Furthermore, changes in fluvial transport length can be more-or-less neglected, as this is regarded to be near constant through time. Water content is a parameter strongly linked to the bulk density parameter, as water fills the pores and expresses the porosity of the sediment (Menounos, 1997).

By using regression models explaining the relationship between ELA and DBD values it is possible to construct coherent ELA reconstructions based on the DBD values. Grain-size analyses can be used to validate the signal (e.g. sorting anomalies).

#### Paraglacial reworking of glacial sediments

A common problem in interpreting proglacial lake sediments is paraglacial reworking of sediments incorporated in 'true' glacially derived sediments. The traditional definition of the term "paraglacial" by Church and Ryder (1972) is "nonglacial processes that are directly conditioned by glaciation". It is widely used to describe the reworking of glacial deposits by rivers and slope processes after the withdrawal of glacier ice (Ballantyne and Benn, 1994). The definition is set into a wider context by Ballantyne (2002) who reformulated the term to "nonglacial earth-surface processes, sediment accumulations, landforms, landsystems and landscapes that are directly conditioned by glaciation and deglaciation". In proglacial lakes paraglacial



**Figure 9** (A) Regression between bulk density values and time resolution (yr/cm). The coefficient of determination (R-squared) shows high predictability due to sedimentation rate through the bulk values. (B) Modelled age-depth based on average bulk values between the radiocarbon dates in the compiled lithostratigraphy from Dravladalsvatn (compiled lithology in paper 3). The upper low organic part is best reproduced in the model. The lower part of the figure shows the difference between the modelled age-depth and the age-depth model based on the radiocarbon dates.

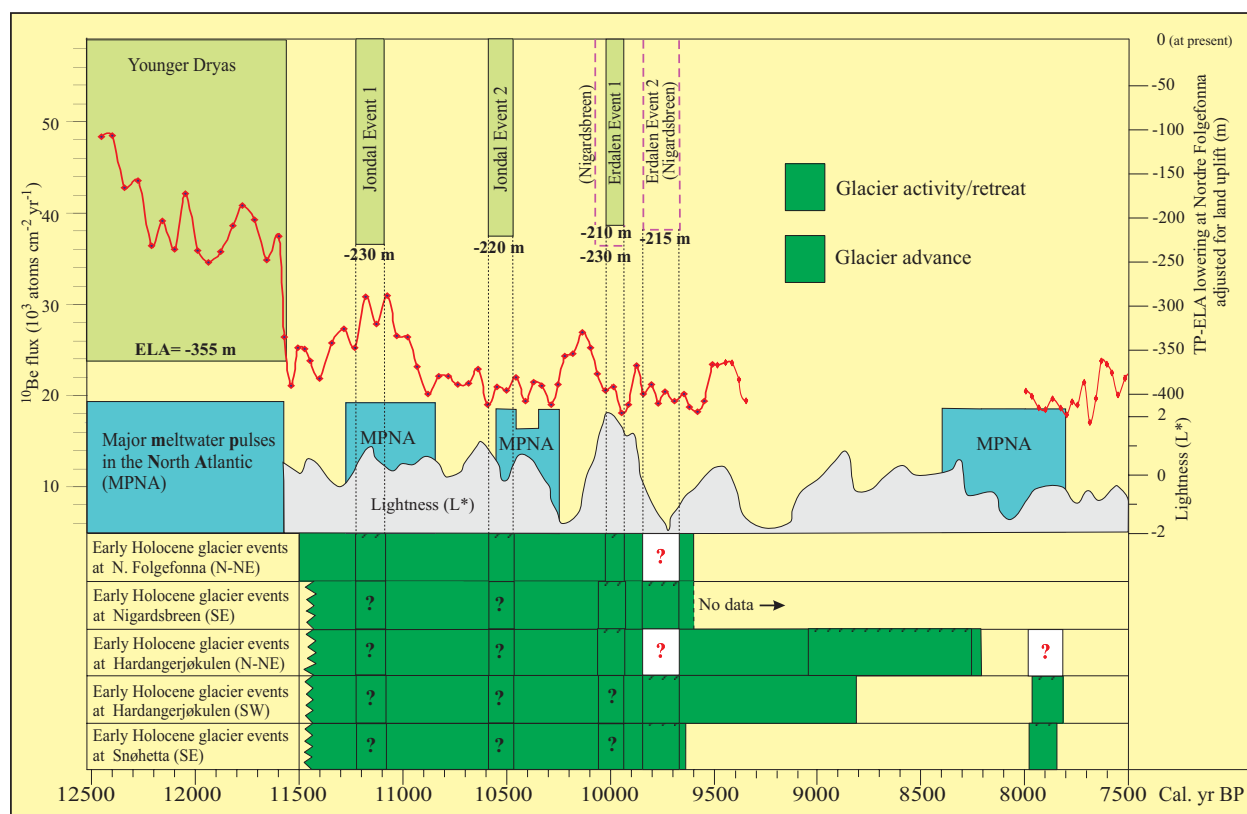
sediments represent “noise” regarding the minerogenic sediments produced directly by the glacier. For reducing the “signal-to-noise-ratio” it is therefore favourable to work in catchments with a sparse cover of superficial deposits.

In the high mountain areas of the Folgefonna Peninsula the potential for paraglacial processes is reduced as the acid Precambrian bedrock is weathering resistant and hence consists of only restricted areas with potential sediments subject to paraglacial activity. This is illustrated in Figure 8 where the amount of silt input to Vetlavatn is plotted during the Lateglacial and early-Holocene. As seen from Figure 8, the “paraglacial cycle” (Church and Ryder, 1972) is rather short, but enhanced during periods with heavy rainfall as illustrated during the Erdalen Event 2.

In the Lyngen Peninsula the potential for paraglacial activity is larger, as the studied catchment area (paper 4) is covered by large amounts of superficial deposits. This is also seen in lake sediments from Lake Aspvatn, where several flooding events are seen as individual minerogenic

layers with similar physical properties as glacially derived sediments. This problem has to some extent been solved by the use of grain-size analysis. The grain-size variations in lake sediments can be linked to the energy in the hydrological system, especially catastrophic events such as river flood and avalanches. We have used the relationship between ‘sorting’ and ‘mean’, two parameters derived through statistical analyses of the derived grain-size data. ‘Mean’ and ‘sorting’ are both sensitive to abrupt hydrological energy changes. As most of the paraglacial reworking is related to flooding events, this approach can be used to distinguish between paraglacial sediments and direct glacier-derived sediments.

However, it is always debatable whether minerogenic sedimentation in proglacial lakes can be regarded as a measure of actual glacier size or if other catchment processes influence the deposition. Using grain-size analyses combined with comprehensive mapping of actual processes in the catchment area can to some extent solve the problem with paraglacial reworking.



**Figure 10** Glacier events at northern Folgefonna, Hardangerjøkulen, Snøhetta and at Nigardsbreen (Dahl and Nesje, 1996; Dahl et al., 2002) with corresponding ELA lowering estimates to the right, meltwater pulses in the North Atlantic (Clark et al., 2001), lightness (Chapman and Shackleton, 2000) with scale to the right and atmospheric Beryllium fluxes measured in the GRIP ice core (Björck et al., 2001, and references therein) with scale to the left. The timing of the early-Holocene glacier events at northern Folgefonna and at Nigardsbreen are shown in the lower column (Paper 2).

### Age-depth modelling

The study of lake sediments requires sufficient age control to enable comparisons and correlations on local, regional and global scales. In this thesis the age-depth control in core records is adopted through radiocarbon dating of macrofossils and bulk AMS samples when terrestrial macrofossils were not available. Recent comparisons between dated bulk sediment and macrofossil samples from various lakes commonly show marked discrepancies. It has therefore been put effort into extracting terrestrial macrofossils for radiocarbon dating. In some parts of the records presented in the thesis this has been difficult and AMS bulk dates were used. However, on certain sites and under certain conditions, AMS dates on terrestrial plant macrofossils are not more precise than bulk sediment samples. If the bedrock consists of acid bedrock and the organic content is sufficient, we have obtained reliable results from AMS bulk dates compared with levels dated with terrestrial macrofossils. However, at some sites we have also discovered large discrepancies between the two techniques.

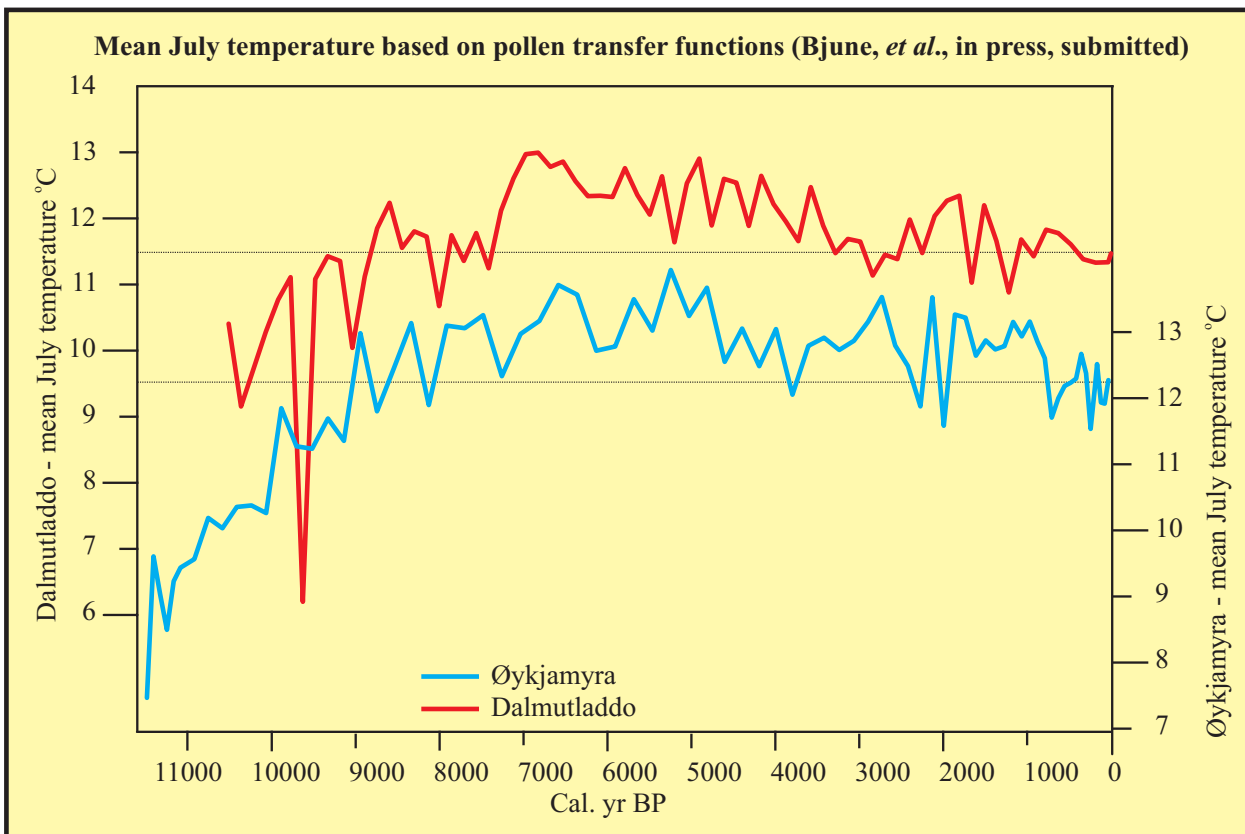
Irregular inflow of minerogenic sediments makes the age-depth modelling complicated in proglacial lakes and in lakes with active catchment processes (e.g. flooding or avalanching). It is therefore necessary to date both above and below sediments that are suggested to be linked to rapid environmental changes. This sampling strategy has

to some extent been followed in this thesis, but is limited by available financial resources for radiocarbon dating.

In lakes with more-or-less stable sedimentation rates, the age-depth modelling is commonly done by using a weighted regression procedure in the framework of generalized additive models (Heegaard *et al.*, 2004). In proglacial lakes, where the sedimentation rates are irregular, the age-depth models using additive models are inappropriate. We have therefore used linear interpolation between each radiocarbon-dated level. By using linear interpolation, age-depth models typically consist of abrupt changes in accumulation rates near the radiocarbon dates. This is sometimes caused by the sampling strategy where radiocarbon dates are commonly taken in areas with lithological changes. Hence, linear interpolation can overestimate the real sedimentological change. In paper 3 we tested the DBD (dry bulk density) values against sedimentation rates, as it is assumed that there should be an accumulation rate signal within the DBD values (Figure 9). The test was done by using simple regression models, and it reproduces some of the same shifts in sedimentation rates as seen from the linear age-depth model.

Using radiocarbon dates is complicated by many possible problems, which are reviewed by Björck and Wohlfarth (2001). The different age-depth models used in this thesis is discussed in each paper.





**Figure 11** Inferred  $T_{jul}$  based on pollen transfer functions from Dalmutladdo and Øykjamyra shown in absolute temperatures °C (Bjune *et al.*, in press, submitted, respectively). Black line denotes present day climate at study site.

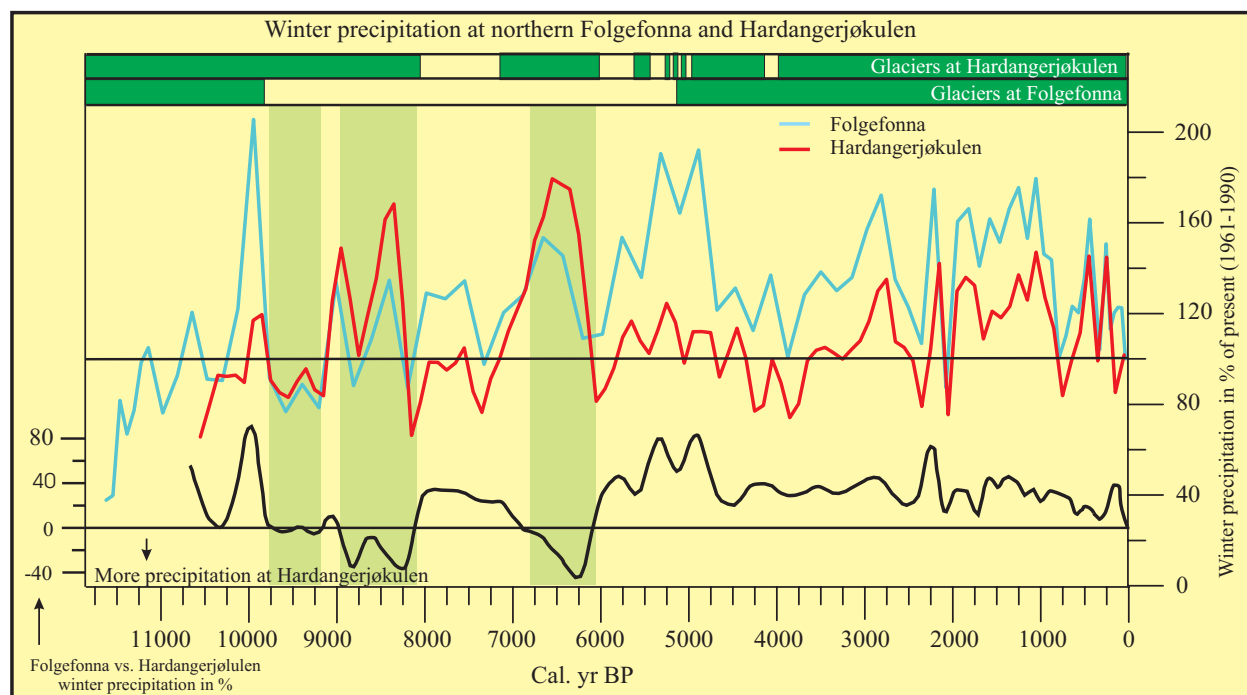
### Climatic implications of the event chronology and the winter precipitation reconstructions

Abrupt climate events have been recorded in a number of climate proxies and archives in the North Atlantic region during the last glacial period and the early-Holocene (Dansgaard *et al.*, 1993; Bond *et al.*, 1997; Adams *et al.*, 1999). This is also seen from the reconstructions presented in this thesis, and some attainable linkages of broader regional importance and also possible linkages to forcing mechanisms seen from proxies around the North Atlantic are therefore discussed. The Lateglacial and early-Holocene records from southern Norway are discussed in the context of thermohaline circulation proxies, solar radiation (as seen through  $^{10}\text{Be}$  isotopes) and meltwater pulses from the Laurentide and Scandinavian ice sheets (Figure 10). Finally, based on the new temperature reconstructions adopted through the NORPEC project (Figure 11), we have recalculated and discussed the pattern of winter precipitation along transects from west to east, and from south to north. The comparisons are based on ELA reconstructions presented in this thesis and one previously published record from Hardangerjøkulen (Dahl and Nesje, 1996).

#### Lateglacial and early-Holocene glacial events - ocean-atmosphere interactions

The INTIMATE group suggested that the GRIP

Greenland ice core should form the stratotype for the last termination (Björck *et al.*, 1998). These data have been used to calibrate different proxies from all over the Northern Hemisphere. In this thesis we have investigated glacier fluctuations during the transition from glacial to interglacial both in southern and northern Norway. Until recently there has been a lack of data examining the first 1500 yr after the Younger Dryas in Scandinavian glacier reconstructions. A common problem is long lasting remnants of the Scandinavian Ice Sheet (SIS) limiting the extent of the lacustrine records backwards in time. To some extent however, this is solved through the record from Folgefonna combined with the glacial records from Jostedalbreen and Hardangerjøkulen, giving timing and magnitude of the events from the Younger Dryas to the mid-Holocene (Figure 10). In northern Norway the record from Strupskardet covers this time span as the studied area lies beyond the limit for SIS during most of the Lateglacial. To avoid long lasting ice sheets we have studied glaciers situated in two peninsulas (Figures 2 and 3), as the surrounding fjords act as draining channels for the SIS, and thereby also isolates the climatically controlled glaciers during the early part of the deglaciation. This is also seen in a study from the Ålfotbreen area, which had an isolated ice cap during the Younger Dryas (Sønstegaard *et al.*, 1999). This assumption is demonstrated through studies of sea-level fluctuations (Helle *et al.*, 1997; Helle, 2004) where it is suggested that the Hardangerfjord was



**Figure 12** Holocene winter precipitation variations at the present ELA compared to present values (1961-1990 = 100%) at northern Folgefonna and at Hardangerjøkulen (Dahl and Nesje, 1996). The lower figure shows the difference between the two records.

deglaciated as early as 12.5 ka  $^{14}\text{C}$  BP. The early deglaciation of the Hardangerfjord is disputed (Mangerud, 2000), but data in this thesis show that the high-altitude plateau of the Folgefonna Peninsula was isolated from the SIS during the Younger Dryas (Paper 2).

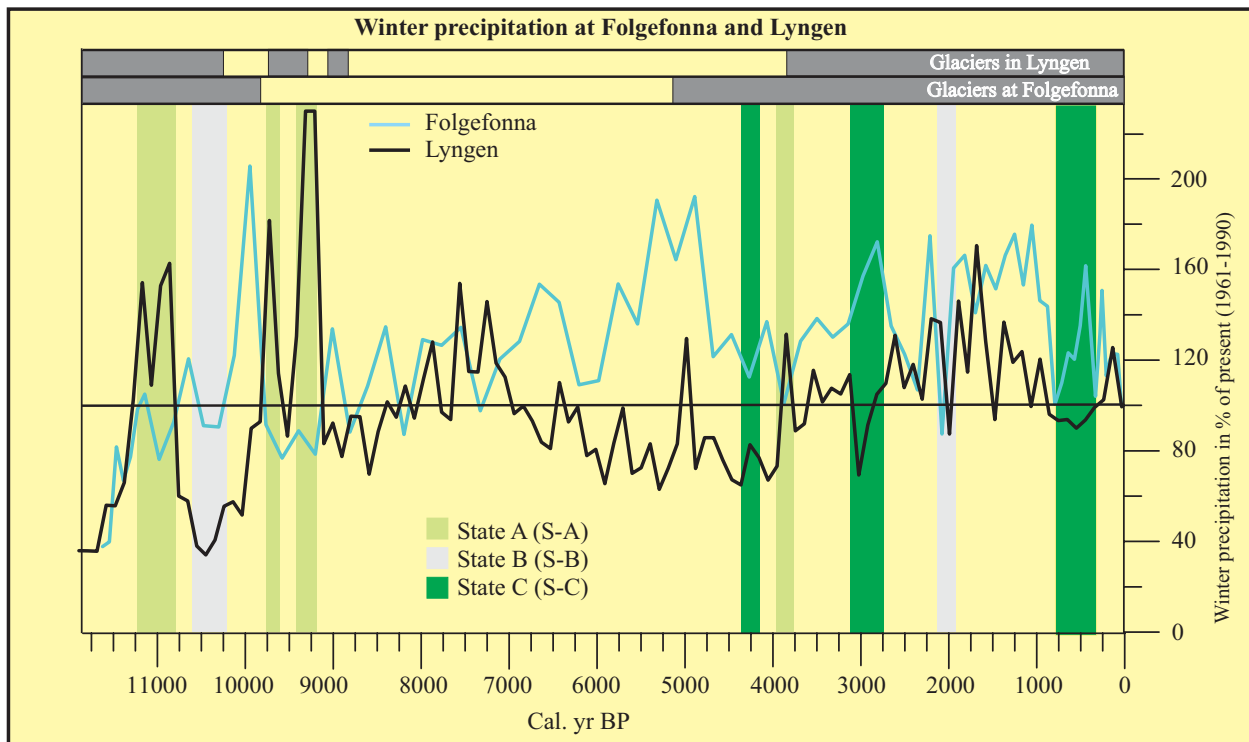
The glacial event chronology from southern Norway during the early-Holocene starts with the “Jondal Event 1” (11,150-11,050 cal. yr BP) followed by “Jondal Event 2” (10,550-10,450 cal. yr BP), “Erdalen Event 1” (10,050-9950 cal. yr BP), “Erdalen Event 2” (9850-9650 cal. yr BP), “Finse Event1” (9000-8400 cal. yr BP) and finally “Finse Event2” (8000-7800 cal. yr BP). The naming of the events is done according to the place where the glacial readvances was first recognized as climate induced events. Many of the events are later reproduced from other sites and in other archives (Eikeland, 1991; Hald and Hagen, 1998; Klitgaard-Kristensen et al., 1998; Matthews et al., 2000; Nesje et al., 2000a; Nesje and Dahl, 2001; Nesje et al., 2001; Dahl et al., 2002; Seierstad et al., 2002).

The first glacial readvance (Jondal Event1) is suggested to be synchronous with the Pre-boreal Cooling (PBO) (Björck et al., 1997) where the causal explanation is lowered temperatures with a possible linkage to solar radiation as indicated through increased  $^{10}\text{Be}$  in ice cores from the Greenland Ice Sheet (GIS). The event is also confirmed in a pollen-based temperature record from the Folgefonna Peninsula (Bjune et al., Submitted) (Figure 11). The remaining early-Holocene glacier readvances are suggested to mainly be the result of increased winter precipitation. Both Jondal Events terminated abruptly with a possible link to the shutdown of meltwater from the Laurentide Ice Sheet (LIS) (Clark et al., 2001). It is suggested that freshwater releases

to the North Atlantic would give reduced production of North Atlantic Deep Water (NADW), which is supposed to give colder SST temperatures in the North Atlantic, and hence, also lower winter precipitation at the west coast of Norway (Chapman and Shackleton, 2000; Clark et al., 2001; Feddersen, 2003; Bakke et al., Submitted). The same pattern is seen during the Finse Event. Common for all the glacier readvances during the early-Holocene is that they appear during periods with relatively high overturning and hence high production of NADW in the North Atlantic (Chapman and Shackleton, 2000). All the studied glaciers are maritime or semi-continental and the decrease in winter precipitation may have led to precipitation starvation at the studied glaciers. In paper 2 we suggest that the ‘Jondal Event 2’ is the first climatic event where the modern circulation was established over the North Atlantic region with strong westerlies and precipitation derived from a relatively warm North Atlantic Ocean.

### Winter precipitation along a east west transect in southern Norway

Using winter precipitation as a proxy for atmospheric circulation, it is essential with reconstructions along transects in order to examine possible gradients and shifts in snow-bearing wind directions. We have therefore compared the reconstructed winter precipitation records from northern Folgefonna with the record from Hardangerjøkulen (Dahl and Nesje, 1996). As temperature is more-or-less regional in southern Norway, we have used the same temperature reconstruction (Figure 11) for both Hardangerjøkulen and northern Folgefonna when calculating the former winter precipitation.



**Figure 13** Holocene winter precipitation variations at the present ELA compared to present values (1961-1990 = 100%) at northern Folgefonna and in Lyngen. Three different modes are identified during the Holocene.

The overall pattern seen in Figure 12 is that northern Folgefonna has on average more precipitation than Hardangerjøkulen. This is as expected since the prevailing snow-bearing wind direction in southern Norway is from southwest at present. Three periods during early to mid-Holocene (9800 – 9200, 9000 – 8100 and 6800 – 6000 cal. yr BP) show higher winter precipitation at Hardangerjøkulen than at Folgefonna. This is suggested to have been caused by a somewhat changed atmospheric circulation pattern when the prevailing snow-bearing winds may have changed to a more south-easterly direction. Seen in modern analogues this is possible when the atmospheric polar front is enhanced and move south of 60 N°. The precipitation follows the anti-cyclonal low-pressure fields to the south of Skagerrak and gives humid air at the eastern side of the mountains that normally are in the precipitation shadow.

#### Winter precipitation in a south-north transect along the west coast of Norway

By combining two winter precipitation records, one from the Lyngen Peninsula and one from northern Folgefonna (Figure 13), it is suggested that some periods with changed atmospheric circulation over the North Atlantic region occurred during the Holocene. Ballantyne (1990) suggested that the atmospheric polar front was lying further to the north during the end of the 'Little Ice Age' causing the late readvance in northern Norway. It is also shown from instrumental records that the precipitation gradient from southern to northern Norway can be reversed (Figure 13).

During the early- to mid-Holocene it is indicated that the climate was in an unstable mode with large fluctuations in winter precipitation. This is seen in the compared winter precipitation records, but the record from Lyngen also suggests four periods with distinct increases in winter precipitation. These events are out of phase with the early-Holocene precipitation events in southern Norway. The North Atlantic region underwent a series of abrupt climatic oscillations when the Northern Hemisphere ice sheets retreated during the last glacial termination. Evidence for these oscillations is recorded in several palaeoclimatic archives (e.g. Nesje *et al.*, 2004). However, little is known about the configuration of the atmospheric circulation pattern during this time span. Synthesis of winter precipitation records can be used in up-scaled climate models and hence give valuable information regarding former circulation patterns.

Three different climate states are suggested during the Holocene when comparing the two winter precipitation records from Lyngen and Folgefonna. State A (S-A) is situations where it was relatively high precipitation in the north and dry in the south, which is suggested to reflect periods with northward drift of the polar front (Ballantyne, 1990). It is important to note that the comparison is done with respect to present precipitation (1961-90 normal = 100%) values. Variations are therefore not reflecting the absolute amount of precipitation, but compared to the 1961-90 normal.

State B (S-B) is situations where the relative precipitation pattern showed the same trends at both sites. This is suggested to reflect periods with general

higher or lower influence of the westerlies.

State C (S-C) is situations where the winter precipitation at Folgefonna was significantly higher than the record from Lyngen. This is suggested to reflect periods when the polar front was lying over southern Norway. During these periods, prevailing westerlies hit southern Norway with a reduced transport of humid air northwards.

The above discussion highlights future perspectives on the importance of reliable winter precipitation estimates from a larger geographical area. Fundamental prerequisites for comparable data sets are constrained summer temperature reconstructions and TP-ELA reconstructions based on independent age-depth models.

## Presentations of papers

### Paper 1

Dahl, S.O, Bakke, J., Lie, Ø., and Nesje, A. 2003: Reconstruction of former glacier equilibrium-line altitudes based on proglacial sites: an evaluation of approaches and selection of sites. *Quaternary Science Reviews* 22, 275-287.

Various approaches are used to record variations in glacier activity and equilibrium-line altitudes (ELAs) based on proglacial sites (lacustrine and terrestrial). These approaches are based on a conceptual model of glacier-meltwater induced sedimentation in which the minerogenic (nonorganic) component of the sediments is related to the occurrence of a glacier in the catchment. The principal coupling to former glacier activity and ELAs is common for these approaches. However, different methods and techniques may complement each other, and both possibilities and limitations are demonstrated. Site selection for reconstructing variations in glacier activity/ELAs is evaluated and critical factors are discussed. Rerouting of glacier meltwater streams across local watersheds in combination with proglacial sites gives a distinct on/off signal for former glacier activity/ELAs. Together with representative lateral moraines of known age, local watersheds are important for calibrating reconstructed glacier activity/ELAs based on a chain of proglacial lakes. Based on the 'modern analogue principle', various proxies can record whenever glaciers existed in a catchment. In a chain of proglacial lakes with different sensitivity to record variations in glacier activity/ELAs, these proxies can be calibrated against independent records. For one-site approaches, however, variations in glacier activity/ELAs depend on the interpretation and sensitivity of the proxies used.

### Paper 2

Bakke, J., Dahl, S.O., and Nesje, A. Lateglacial and early-Holocene palaeoclimatic implications based on reconstructed glacier fluctuations and equilibrium-line altitudes at northern Folgefonna, Hardanger, western Norway.

Submitted to *Journal of Quaternary Science*.

**Abstract:** Northern Folgefonna (c. 23 km<sup>2</sup>), is a nearly circular maritime ice cap located on the Folgefonna Peninsula in Hardanger, western Norway. By combining marginal moraines and AMS radiocarbon dated glacier-meltwater induced sediments in proglacial lakes draining northern Folgefonna, a continuous highresolution record of variations in glacier size and equilibrium-line altitudes (ELAs) during the Lateglacial and early-Holocene has been obtained. After the termination of the Younger Dryas (c. 11,500 cal. BP), a short-lived (150-200 years) climatically induced glacier readvance termed 'Jondal Event 1' occurred within the 'Preboreal Oscillation' (PBO) c. 11,200 cal. BP. Bracketed to 10,550-10,350 cal. BP, a second glacier readvance is named the 'Jondal Event 2'. A third readvance occurred about 10,000 cal. BP and corresponds to the 'Erdalen Event 1' recorded at Jostedalsbreen. An exponential relationship between mean solid winter precipitation and ablation-season temperature at the ELA of Norwegian glaciers makes it possible to reconstruct former variations in winter precipitation if the corresponding ELA is known and combined with an independent proxy for summer temperature. Compared to at present, the Younger Dryas was much colder and drier, the 'Jondal Event 1'/PBO was colder and somewhat drier, the 'Jondal Event 2' was much wetter, whereas the 'Erdalen Event 1' started as rather dry and terminated as somewhat wetter. Variations in glacier magnitude/ELAs and corresponding palaeoclimatic reconstructions at northern Folgefonna suggest that low-altitude cirque glaciers (lowest altitude of marginal moraines 290 m asl.) in the area existed for the last time during the Younger Dryas. These low-altitude cirque glaciers of suggested Younger Dryas age do not fit into the previous reconstructions of the Younger Dryas ice sheet in Hordaland.



### Paper 3

Bakke, J., Lie, Ø., Nesje, A., Dahl, S. O., and Paasche, Ø. A high-resolution glacial reconstruction based on physical sediment parameters from proglacial lakes at Folgefonna, western Norway.

*Submitted to The Holocene.*

**Abstract:** The maritime plateau glacier northern Folgefonna in western Norway has a short (subdecadal) response time to climate shifts, and is therefore well suited for reconstructing high-resolution glacier fluctuations. The reconstruction is based upon physical sediment parameters in two proglacial lakes and a peat bog north of the ice cap. The record of glacier variations has been transferred into an equilibriumline altitude (ELA) variation curve. Glaciers respond primarily to changes in summer temperature and winter precipitation. At present there is a high correlation between the North Atlantic Oscillation (NAO) index and measured (since the early 1960s) net mass-balance on maritime glaciers in western Norway ( $r = \sim 0.8$ ). Reconstructed glacier variations from the maritime western Norway can therefore be indicative of NAO-like weather modes during the Holocene. The early phase of mid-Holocene glacier growth was characterized by gradual expansion leading up to the first Subatlantic glacial event at 2300 cal. yr BP. The climate during the last 2200 years has been favourable for glacier growth at Folgefonna. High-amplitude shifts in ELA are interpreted as a result of unstable modes of the westerlies at the west coast of Norway, with significant changes in winter precipitation. Plotting the bulk density curve against the modelled glacier net mass-balance shows a remarkably similar pattern, where the maximum sediment yield was delayed by  $\sim 10$  years with respect to glacier net mass-balance. We here present a new method of reconstructing glacier variations in areas with low organic production. The approach is highly relevant in highalpine and arctic regions where high-resolution reconstructions of former glacier variations are sparse.

### Paper 4

Bakke, J., Dahl, S.O., Paasche, Ø., Løvlie, R., and Nesje, A. Late Weichselian and Holocene glacier fluctuations, equilibrium-line altitudes and palaeoclimate in Lyngen, northern Norway: a high-resolution composite chronology based on lacustrine and morpho-stratigraphical evidence.

*Submitted to Quaternary Science Reviews.*

**Abstract:** Based on lacustrine and morpho-stratigraphical evidence from Lyngen in Troms, northern Norway, thirteen marginal moraines have been mapped and dated in front of Lenangsbreen in Strupskardet. Moraines M1-M3 and M8-M9 are

suggested originally to be formed as ice-cored moraines, whereas M4-M7 and M10-M13 are suggested to be push and melt-out moraines. A chronological framework, taking into account a combination of relict periglacial features, former shorelines and related glacier-meltwater channels, lichenometry, and AMS radiocarbon dated lacustrine sediments spanning the last 20,000 cal. yr BP, has been established. The proglacial lake Aspvatnet was isolated from the sea c. 10,300 cal. yr BP, and the lacustrine sediments have been investigated by use of LOI at 550/950°C, magnetic susceptibility, water content, wet- and dry bulk density (DBD), and the magnetic parameters anhysteretic remanent magnetisation (ARM) and saturation remanent magnetisation (SIRM). There is in general good agreement between physical sediment parameters and magnetic parameters. DBD, a combination of medium and fine silt and the two statistical parameters 'sorting' and 'mean' have been used to construct a high-resolution glacier-fluctuation curve for the last 3800 cal. yr BP. Based on an accumulation-area ratio (AAR) of 0.6 and an adapted 'Little Ice Age ratio' (LR) approach, a continuous temperature-precipitation-wind equilibrium-line altitude (TPW-ELA) curve for the last 20,000 cal. yr BP have been constructed. Using an established exponential relationship between mean ablation-season temperature and mean annual solid precipitation at the ELA of Norwegian glaciers, variations in mean winter precipitation as snow are quantified using an independent proxy for summer temperature. Mean annual winter precipitation varied from 500 to 5000 mm water equivalents, and on average, Holocene estimates is c. 50% higher than similar figures from the Lateglacial. The two driest periods occurred during Heinrich events 1 (H1) (17,500-16,500) and 0 (H0) (13,000-12,200), whereas freshwater pulses to the North Atlantic had apparently no systematic impact on mean winter precipitation. Based on the winter precipitation curve from Lyngen, the atmospheric circulation responded to the SST lowering associated with H1 and H0 events in Lyngen with formation of talus-derived rock glaciers at sea level.

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## Part II

### Paper 1

Dahl, S. O., Bakke, J., Lie, Ø. and Nesje, A. 2003. Reconstruction of former glacier equilibrium-line altitudes based on proglacial sites: an evaluation of approaches and selection of sites. *Quaternary Science Reviews* 22 , 275-287.







PERGAMON

Quaternary Science Reviews 22 (2003) 275–287



# Reconstruction of former glacier equilibrium-line altitudes based on proglacial sites: an evaluation of approaches and selection of sites

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Received 1 February 2002; accepted 24 July 2002

## Abstract

Various approaches are used to record variations in glacier activity and equilibrium-line altitudes (ELAs) based on proglacial sites (lacustrine and terrestrial). These approaches are based on a conceptual model of glacier-meltwater induced sedimentation in which the minerogenic (nonorganic) component of the sediments is related to the occurrence of a glacier in the catchment. The principal coupling to former glacier activity and ELAs is common for these approaches. However, different methods and techniques may complement each other, and both possibilities and limitations are demonstrated. Site selection for reconstructing variations in glacier activity/ELAs is evaluated and critical factors are discussed. Rerouting of glacier meltwater streams across local watersheds in combination with proglacial sites gives a distinct on/off signal for former glacier activity/ELAs. Together with representative lateral moraines of known age, local watersheds are important for calibrating reconstructed glacier activity/ELAs based on a chain of proglacial lakes. Based on the 'modern analogue principle', various proxies can record whenever glaciers existed in a catchment. In a chain of proglacial lakes with different sensitivity to record variations in glacier activity/ELAs, these proxies can be calibrated against independent records. For one-site approaches, however, variations in glacier activity/ELAs depend on the interpretation and sensitivity of the proxies used.

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## 1. Introduction

Beyond the time span covered by observed mass-balance measurements and historical records, little was known until recently about continuous variations in equilibrium-line altitudes (ELAs) on Norwegian glaciers. Former ELAs based on the maximum altitude of lateral moraines only reflect shorter periods when the glaciers attained steady state in advanced positions beyond later glacier advances. With the exception of some few glaciers in the continental eastern Jotunheimen (e.g. Matthews and Shakesby, 1984) and at the maritime Folgefonna (Bakke et al., 2000), the marginal moraines formed by glacier advances after the Erdalen Event (Preboreal/Boreal transition; ca 10,000 cal. BP) were erased by the Little Ice Age advance (ca AD 1750) in southern Norway (Matthews, 1991; Dahl and Nesje,

1994). Hence, variations in the ELA based on lateral moraines cannot be used for most of the Holocene in this region.

To avoid this problem, lacustrine and terrestrial deposits in proglacial lakes and bogs beyond the maximum Little Ice Age glacier position have been used to reconstruct periods of former glacier activity within a catchment. Erosion along the glacier sole produces rock floor consisting of clay and silt, and this is transported downstream as suspended material in proglacial meltwater streams (Fig. 1). In contrast to normally transparent nonglacial streams, turbid proglacial meltwater streams typically deposit accumulations of bluish-grey sandy and/or clayey silt that can be used as a signature for the existence of glaciers within the catchment (e.g. Karlén, 1976). With no glaciers in the catchment, lacustrine sediments with a higher organic content (gyttja) dominate in the proglacial lakes, while peat or gyttja accumulate at the terrestrial sites.

Changes in mean ablation-season temperature (1 May–30 September) and winter precipitation (1 October–30

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Fig. 1. Photo showing the sharp contrast indicated by red arrow between a turbid glacier meltwater stream from Hardangerjøkulen and a clear nonglacial stream just east of Finse, central southern Norway. The erosion along the glacier sole produces rock flour consisting of clay and silt, and transported downstream as suspended material this gives proglacial meltwater streams their characteristic green-grey colour. Turbid proglacial meltwater streams typically deposit bluish-grey sandy and/or clayey silt that can be used as a signature for whenever upstream glaciers existed. (Photo: S. O. Dahl).

April) are reflected as variations in the ELA (cf. Andrews, 1975). Based on the mean ablation-season temperature and winter precipitation at the ELA of Norwegian glaciers in different climatic regimes, an exponential relationship has been demonstrated (Dahl et al., 1997, and references therein). This relationship implies that if either the winter precipitation or the ablation-season temperature at the ELA is known, the other factor can be calculated. It also implies that if the former ELA is known, it is possible to calculate how the other parameter has fluctuated (Dahl and Nesje, 1996). Biological proxies (e.g. beetles, chironomids, mites, pine-tree limits, plant-macro fossils, pollen, etc.) from sites not influenced by glacier meltwater or streams from 'permanent' snowfields can be used as independent proxies for summer temperature, while the potential for reconstructing continuous variations in the ELA depends on the number and type of ELA-related proglacial sites at each glacier.

In recent years, a number of curves showing Holocene ELA fluctuations have been published from various

regions of southern Norway. Except for the tentative curve of Liestøl (1969), both relative reconstructions (Nesje and Dahl, 1991; Nesje and Kvamme, 1991; Nesje et al., 1991; Matthews and Karlén, 1992) and absolute reconstructions (Dahl and Nesje, 1996; Nesje et al., 2001) have been primarily based on proglacial sites. In the relative reconstructions of Nesje and Kvamme (1991) and Nesje et al. (1991) a combination of proglacial sites and independent palynological data was used, while the ELA curve of Matthews and Karlén (1992) was based on proglacial sites reflecting glaciers at different altitudes. Dahl and Nesje (1996) used four proglacial sites with different sensitivity to variations in glacier size (GS) in the absolute reconstruction of former ELAs at Hardangerjøkulen, while the absolute reconstruction of former ELAs from the Jostedalbreen region (Nesje et al., 2001) took into account various sediment parameters (weight loss-on-ignition, grain-size analyses, etc.) and sites related to other glaciers in the area to overcome the problem(s) of a one-site approach.



The objective of this paper is to evaluate approaches and site selection when using proglacial sites to reconstruct former glacier activity/ELAs, and to discuss the general principles behind the calibration and interpretation of these field- and laboratory-based ‘empirical’ relationships.

## 2. Methods reflecting glacier activity

The physical processes in glacial sedimentary environments are often complex (e.g. Ashley et al., 1985). However, a number of methods and techniques have been used to record glacier activity based on proglacial lacustrine and terrestrial sites. The methods are based on a conceptual model of glacier-meltwater induced sedimentation in which the minerogenic (nonorganic) component of the sediments is related to the occurrence of a glacier in the catchment (e.g. Karlén, 1981; Leonard, 1985). The organic component depends on many factors, including bedrock lithology and vegetation cover, local climate (temperature, precipitation, wind), size and aspect of the catchment and the lake, water depth and temperature, coverage of superficial sediments in the catchment, colluvial activity related to slope angles around the lake, as well as anthropogenic impact. The minerogenic component in proglacial lakes depends in addition on factors such as transport distance and the number of intervening lakes acting as sediment traps (Smith, 1978). In most lakes the organic component is much smaller than the minerogenic component, but the relative importance of these components is to a large extent site- and/or area dependent. Hence, Matthews and Karlén (1992) argued for the use of representative nonglacial lakes as ‘control lakes’ in connection with the use of proglacial lakes in the study of glacier fluctuations. The contrast between these two lake types then clarifies the glacial signal.

Due to the large colour variations (light-/bluish grey to dark brown) with and without a glacier in the catchment, visual description of the layers in a core or section using a Munsell colour chart is a useful first approach. This should be supported or supplemented by various field and laboratory methods and techniques:

- Mineral magnetic susceptibility commonly reflects the concentration of magnetic minerals (e.g. Thompson and Oldfield, 1986), and may be used as an indicator of erosion and transport of clastic sediments which can be linked to glacier activity (e.g. Snowball, 1993; Leemann and Niessen, 1994; Snowball and Sandgren, 1996).
- X-ray diffraction analyses with low-density values may indicate high sediment density and hence increasing glacier activity (e.g. Karlén and Matthews,

1992; Matthews and Karlén, 1992; Leemann and Niessen, 1994; Souch, 1994).

- Weight loss-on-ignition (LOI) estimates the organic content of lacustrine sediments (Dean, 1974; see Heiri et al., 2001, for laboratory procedures), and can together with derived parameters (minerogenic material, dry weight, water content, etc.) be linked to glacier variations (e.g. Karlén, 1976; Leonard, 1986a, b; Nesje et al., 1994, 2000, 2001; Souch, 1994; Snowball and Sandgren, 1996; Menounos, 1997; Matthews et al., 2000).
- Grain-size variations of especially clay and silt may be linked to glacier fluctuations (e.g. Leemann and Niessen, 1994; Matthews et al., 2000; Nesje et al., 2001; see Blott and Pye, 2001, for the statistical treatment of grain size distribution).
- Occurrence and thickness of clastic varves reflect variations in glacier activity (e.g. Leemann and Niessen, 1994; Leonard, 1997).
- Bulk density is the ratio of the mass of dry (wet) solids to the bulk volume of the sediment, and may be used to record former glacier variations/ELAs (e.g. Leonard, 1997; Menounos, 1997).
- Lithological changes in the recorded sequence or core may reflect the occurrence of glaciers in the catchment and/or variations in glacier extent (e.g. Svendsen and Mangerud, 1997; Matthews et al., 2000).

The principal coupling to former glacier activity/ELAs is common for all these methods and techniques. In Fig. 2 both possibilities and limitations are illustrated in the schematic coupling between weight LOI and former glacier magnitude/ELA, and the glacier events reflect what is typically recorded in a core from a Norwegian proglacial lake with a temperate glacier in the catchment. The proglacial lake is located so far downstream for the glacier that (visible) turbid meltwater only enters the site during the maximum late spring/early summer flood at present. This can also easily be recorded by the bare eye in the top sediments of the core. In periods when the glacier is much smaller or nearly melted away, no visible turbid glacier meltwater enters the proglacial lake and the relative accumulation of organic sediments in the lake is much higher. The occurrence of an active glacier in the catchment cannot be recorded in the sediments by the bare eye, and only various laboratory techniques can detect an input of glacier-induced sediments to the lake. In periods when the glacier is much bigger than at present (in this example, during the early deglaciation), the influence of glacier-induced sedimentation in the proglacial Lake may be so high that variations in glacier magnitude are difficult to detect or separate out by the available laboratory methods. The main problems concerning the interpretation and calibration of these parameters are thus related to when the glacier was

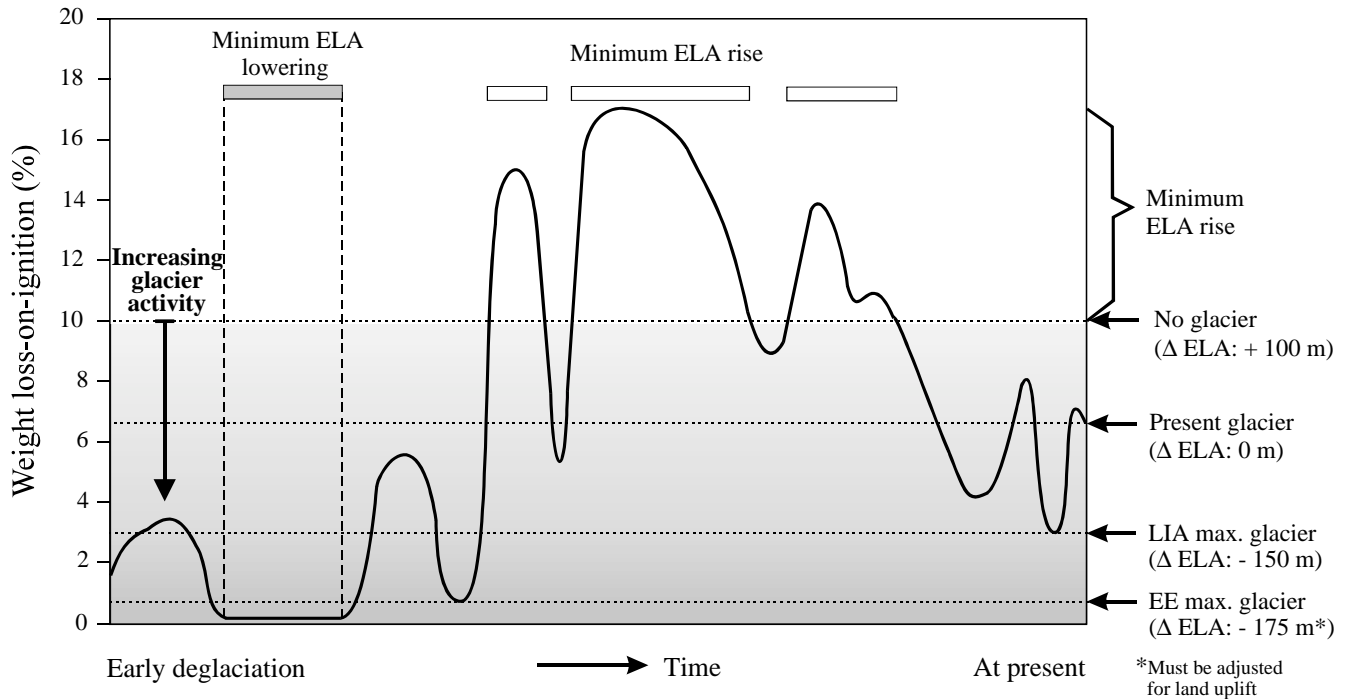


Fig. 2. Schematically illustrated with weight LOI, both possibilities and limitations in the principal coupling between various proxies and variations in glacier activity/ELAs are shown. LIA = Little Ice Age, EE = Erdalen Event. See text for further explanation and discussion.

nearly melted away, or when it was so large that the used proxy lacked sensitivity to record variations in GS. However, the various methods may complement each other.

Both as a link to other palaeoclimatic proxies and as a tool to investigate glacier activity, age control of events are essential. In studies of glacier fluctuations prior to the Little Ice Age maximum several methods have been used to obtain age–depth control:

- By measuring the magnetic declination versus sediment depth, magnetic declination records can be compared with the well-dated master curve of Lake Windermere (Creer and Tucholka, 1983). Assuming synchronous oscillations in the two magnetic declination records this can be used as a dating method (see Snowball, 1993; Leemann and Niessen, 1994, for details).
- The thickness and grain-size distribution in annual clastic glacial varves, if continuous, may represent an annually resolved record of glacier activity. However, the rhythmites in a sequence must be confirmed as real varves in the sense of De Geer (1912) before they can be used for age–depth control (e.g. Leemann and Niessen, 1994; Leonard, 1997).
- Radiocarbon dating tends to be the most important dating method for reconstructing former glacier activity. Recent comparisons between dated bulk sediment and macrofossil samples from various lakes

often show marked discrepancies, and radiocarbon chronologies from lake sediments are often based on AMS-radiocarbon dated terrestrial plant macrofossils (e.g. Barnekow et al., 1998). However, on certain sites and under certain conditions, AMS dates on terrestrial plant macrofossils are not superior to bulk sediment samples (e.g. Gulliksen et al., 1998), and the most reliable chronologies may be obtained not from terrestrial plant macrofossils, but from that part of the sediment fraction (the ‘humic’ NaOH-soluble component), where there is no contamination by older carbon residues (Lowe and Walker, 2000).

Age estimates should be given as calibrated years before present (BP) in accordance with INTCAL98 for radiocarbon dates (Stuiver et al., 1998). As sedimentation rates vary with and without a glacier in the catchment of proglacial lakes, both the initiation and the termination of glacier episodes should be dated. Hence, age–depth control in proglacial lakes relies on linear interpolation within periods with and without a glacier in the catchment.

### 3. Site selection

To find an area with proglacial sites suitable for reconstructing variations in former GS/ELAs, several factors must be evaluated:

- The current glaciated area in the catchment must be ‘appropriate’ for recording the amplitude of glacier variations in the studied time span (e.g. the Holocene). An ideal setting is one in which a single glacier has existed in the catchment throughout the Holocene, but has been so small that only the most sensitive sites have a continuous record of Holocene glacier activity. The area/altitude distribution of the glacier must be taken into account in this evaluation.
- The net mass balance of glaciers (and the ELA) in general depends on the regional distribution of temperature in the ablation season and winter precipitation as snow in the accumulation season. Local redistribution of (dry) snow by wind from exposed surfaces to leeward topographic depressions is in addition of great importance. On ice caps (plateau glaciers) with outlet glaciers in all aspects, the influence of wind can be neglected as the mean ELA will even out the deflation on the windward side and the additional accumulation on the leeward side for any snow-bearing wind direction. Hence, the *local topographic* temperature-precipitation-wind-ELA (TPW-ELA) of cirque glaciers may exist well below the *regional* temperature-precipitation-ELA (TP-ELA) of plateau glaciers (Fig. 3) (Dahl and Nesje, 1992; Dahl et al., 1997). Whether there is a plateau glacier or a cirque glacier in the catchment is thus of importance for the interpretation of the climatic factors influencing the reconstructed glacier activity/ELA.
- A number of downstream proglacial lakes and other sites suitable to document variations in GS/ELAs backwards in time is preferable to a single lake (see below). Without glacier(s) in the catchment the (proglacial) lakes should have a high organic production to differ from periods with glacier-melt-water-induced sedimentation, a requirement often fulfilled by lakes close to the present tree line. Representative nonglacial control lakes (Matthews and Karlén, 1992) should exist in the area.
- Proglacial lakes should be dammed by a rock sill and not by moraines, colluvial fans, rock avalanches, etc. Ideally, lakes with a flat bottom, gentle slopes and with no mixing of the water column to their base should be preferred to minimize post-depositional disturbance of the sediments. Hence, sites close to deltas and other unstable sedimentary environments where turbidites, snow avalanches, etc. commonly occur should be avoided. If possible, the lake bottom surrounding the coring site(s) should be flat in a radius of at least 150 m (Fig. 4).
- The residence time of water in a proglacial lake must be long enough to allow suspended sediments to settle, but short enough to allow some material in suspension to continue and settle in other lakes further downstream.
- To calibrate reconstructed variations in GS/ELAs based on proglacial sites, representative marginal moraines of known age must be found in the catchment (see below).

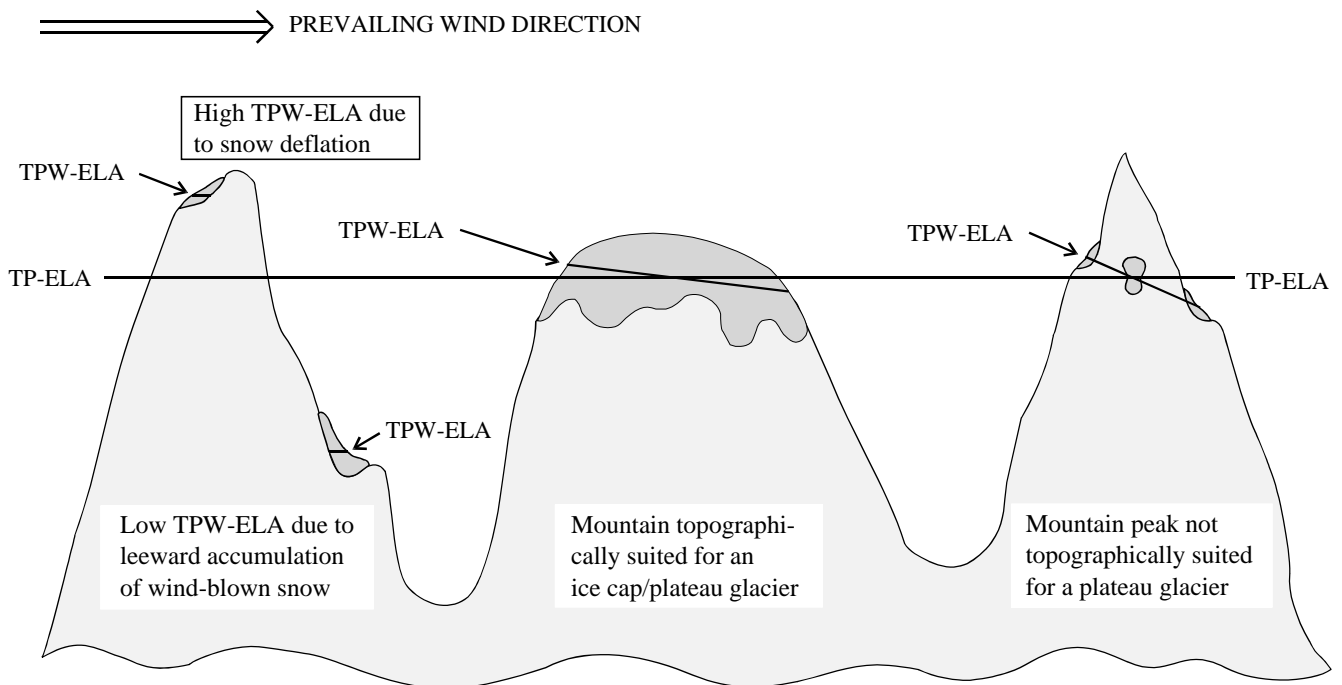


Fig. 3. Schematic examples showing the differences between the regional TP-ELA at plateau glaciers/ice caps and the local topography dependent TPW-ELA at cirque glaciers (Dahl and Nesje, 1992; Dahl et al., 1997).



- Superficial sediments and active geomorphological processes (colluvial activity, floods, etc.) in the catchment may influence lake sedimentation, and must be taken into account.
- In general, ideal proglacial sites turn out to record simple systems where the topographic configuration isolates the glacier meltwater signal through natural filtering in a way that directly reflects GS/ELA.

**4. Reconstruction of former glacier ELAs**

In addition to the maximum elevation of lateral moraines (MELM) (Figs. 4 and 5), the traditional ways to find former ELAs include the median elevation of glaciers (MEG), the toe-to-headwall ratio (THAR), accumulation area ratio (AAR), and the balance ratio method (see Nesje and Dahl, 2000, and references therein). In cases when only sparse remnants of marginal moraines are available, Dahl et al. (2002) introduced a new technique termed the Little Ice Age ratio to estimate the ELA of glacier advances predating the Little Ice Age maximum. In addition to defining the modern ELA of existing glaciers, these techniques are important for calibrating reconstructed ELAs based on proglacial sites.

**4.1. Local watersheds**

If combined with a proglacial lacustrine or terrestrial site, rerouting of glacial meltwater streams across local watersheds may give accurate estimates of former ELAs, if the extent of the corresponding glacier

terminus is known from marginal moraines, historical records, air photographs, etc. (e.g. Dahl and Nesje, 1994; Dahl et al., 2002). Whenever the glacier is in an advanced position beyond the local watershed, glacier-meltwater-induced sediments may be deposited at the proglacial site, while only organic sediments accumulate when the glacier is behind the local watershed. This setting makes it possible to date whenever the glacier and the corresponding ELA are at, or close to, this on-off threshold.

Local watersheds consisting of ‘permanent’ bedrock thresholds are preferred to ensure that this on/off signal has existed throughout the Holocene. Such watersheds (especially inside the Little Ice Age glacier maximum) appear to be near, while rerouting of proglacial meltwater streams caused by ‘temporary’ marginal moraines are more common. Due to the sharp on-off signal related to local watersheds, reconstructed glacier termini are normally very accurate. If the reconstructed glacier front can be linked to a known ELA by AAR, MELM, etc., reconstructed ELA variations related to local watersheds can be used to calibrate ELAs based on downstream proglacial sites (Fig. 4).

**4.2. Chain of proglacial lakes**

Based on data from nine Norwegian glaciers (Roland and Haakensen, 1985) there is a significant correlation ( $r = 0.86$ ) between glacier size/area and calculated sediment transport in proglacial meltwater streams. A similar relationship between the downstream transport distance of glacier-induced sediments in suspension and GS is suggested (Fig. 6), and consequently a chain of

**Site selection related to glacier-melt water induced sedimentation from suspension at proglacial sites**

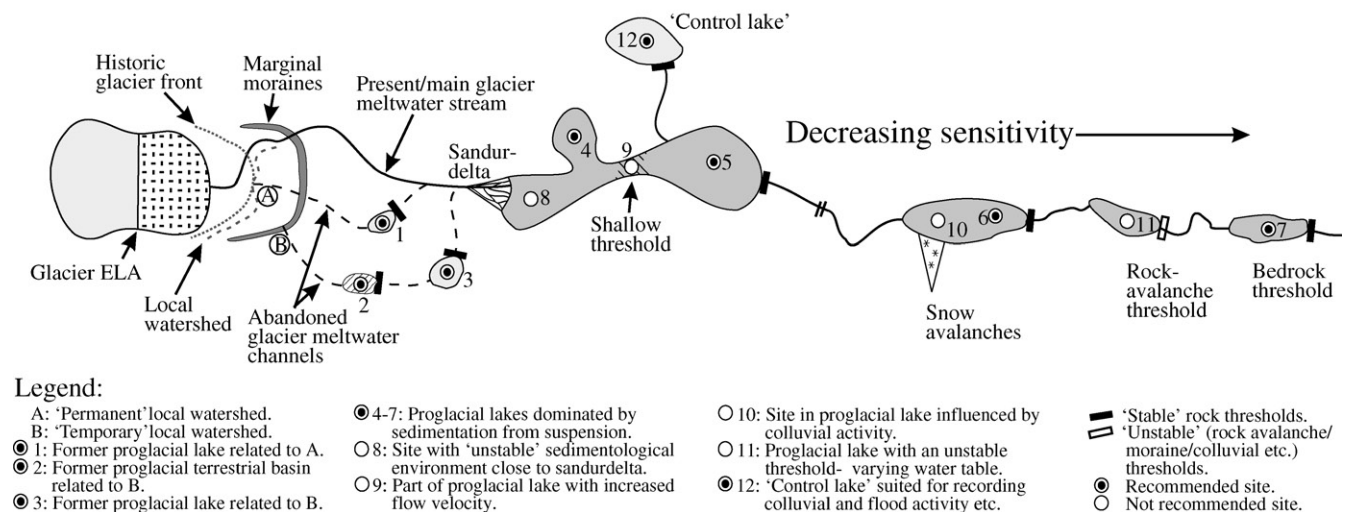
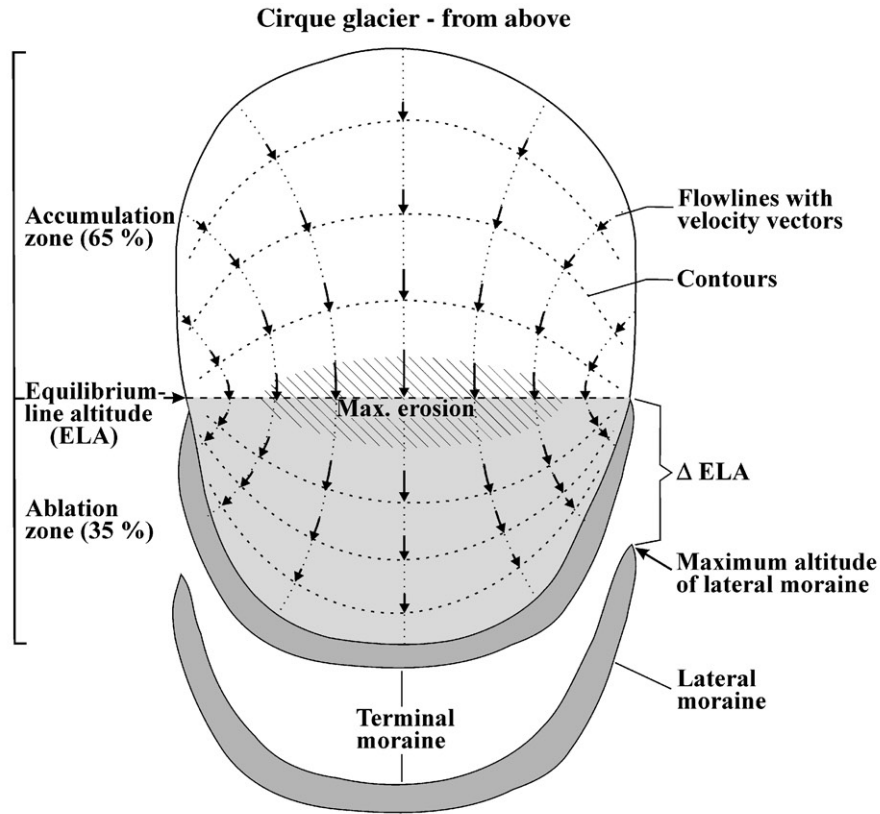


Fig. 4. Potential sites for investigating variations in glacier activity/ELA in a simple catchment with one glacier and a chain of proglacial lakes. Lacustrine and terrestrial sites related to both ‘permanent’ (bedrock) and ‘temporal’ (ice-marginal moraines) local watersheds, control lakes and various settings related to proglacial lakes are shown. Recommended coring sites are marked with partly filled circles, whereas secondary sites close to unstable sedimentary environments (normally not recommended) are marked with open circles. See legend and text for further details.



**Marginal moraines = lateral moraines and terminal moraines**  
**Accumulation-area ratio (AAR) =  $0.65 \pm 0.05$  : 1**

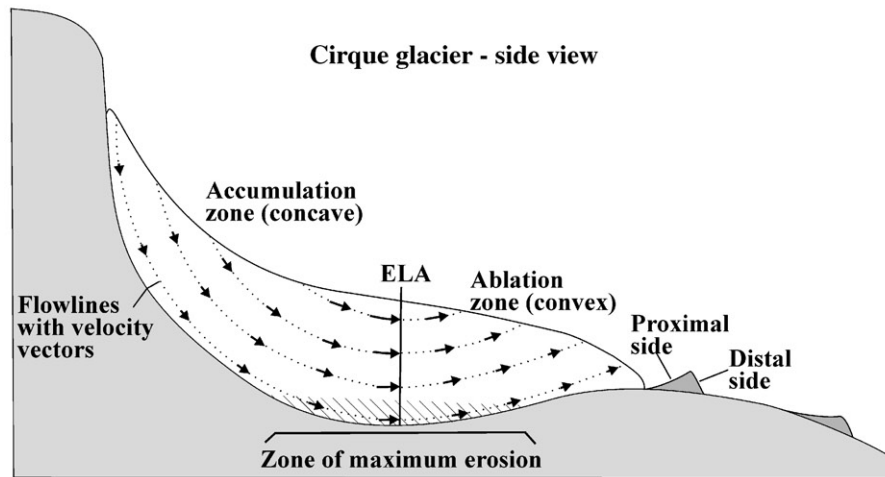


Fig. 5. Flow lines and contours on an idealized cirque glacier seen from above and in side view. Lateral moraines normally do not form above the steady-state ELA, and this can be used to estimate former ELAs. An AAR of 0.65:1 for a cirque glacier in climatic and dynamic steady state is shown. The maximum erosion related to the rotational movement beneath a cirque glacier is closely linked to the ELA, as both flow velocity (illustrated with vectors) and the maximum turnover of ice occur at or close to the steady-state ELA (modified from Lewis, 1949, 1954; McCall, 1952, 1960).

proglacial lakes can be used to record temporary variations in former glacier activity/ELAs. Small glaciers tend only to be recorded at the most sensitive sites, while larger glaciers in addition are recorded at sites further downstream.

Depending on the size and response time, a lowering of the ELA results in a larger glacier when it has obtained climatic steady state. As a larger glacier/lower ELA leads to an increased meltwater discharge, a longer downstream distance of sediments in suspension is

### Schematic relationship between glacier size, glacier ELAs and downstream glacier-meltwater induced sediments

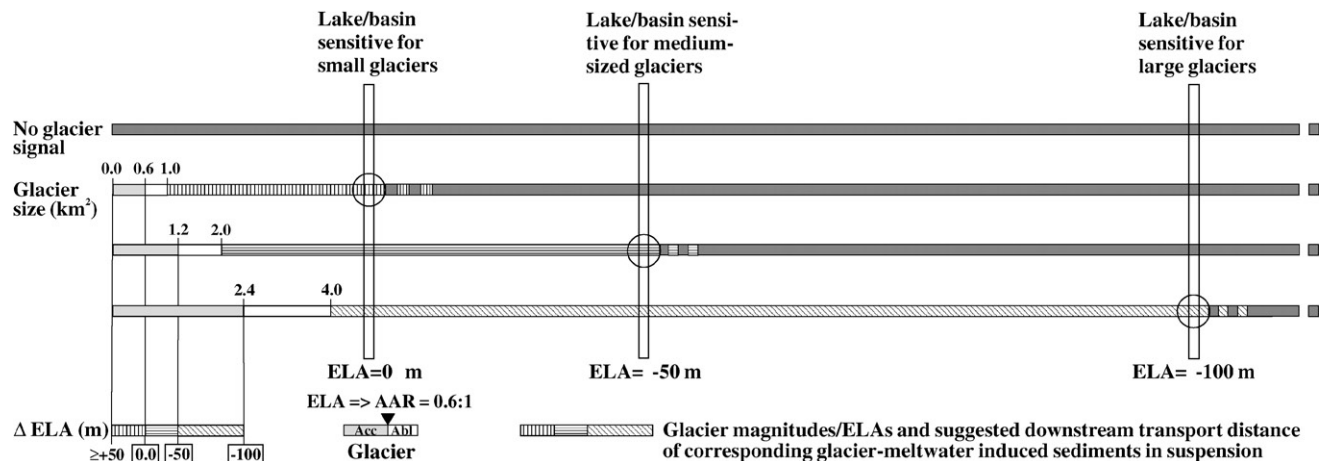


Fig. 6. Schematic relationship between GS ( $\text{km}^2$ ) and SSD (km) in a simple catchment with one glacier and a chain of proglacial lakes. Primarily, the GS/SSD ratio is suggested to be a catchment specific 'fixed' relationship between GS and the downstream distance sediments from suspension can be recorded. In this example a tentative GS/SSD ratio of 1:4 has been used. GS is converted to an ELA estimate using the AAR method. To adjust for catchment dependent factors, however, independent proxies (e.g. lateral moraines, local watersheds, etc.) may be used to calibrate the amplitude of ELA fluctuations. See text for further explanation.

expected. Observations suggest that even a small increase in GS may correspond to an enlarged transport of sediments in suspension for several kilometres downstream (e.g. Dahl and Nesje, 1994). If this GS/sediments in suspension distance ratio (GS/SSD ratio) is known, records of glacier-induced sediments from a chain of proglacial lakes may give sensitive estimates of former variations in glacier magnitude.

The Finse Valley in central southern Norway is a simple catchment with a chain of proglacial lakes completely dominated by meltwater from the northern sector of the ice cap Hardangerjøkulen. Based on modern analogue studies during the ablation season (e.g. Fig. 1), well-dated stratigraphies from both proglacial lakes and basins related to local watersheds (Dahl and Nesje, 1994), and possibilities to establish fixed points for these sites which relate former glacier magnitudes to known ELAs (Dahl and Nesje, 1996), an approximate GS/SSD ratio of 1:4 for this catchment is suggested and used as a tentative example in Fig. 6. Any GS/SSD ratio is suggested to be catchment specific, however, and the ratio depends on a complex interaction between factors like nonglacial tributaries, relief of the river profile, water discharge, residence time of water in the proglacial lakes, etc.

GS given in square kilometres ( $\text{km}^2$ ) is converted to an ELA estimate using the AAR method. To adjust for catchment dependent factors, the amplitude of the glacier/ELA signal in the proglacial lakes can be calibrated by the use of independent ELA observations of known age based on lateral moraines, local watersheds, etc. If the ELA and the corresponding GS/SSD ratio can be established for at least three fixed points

spanning from small to large glaciers, a chain of proglacial lakes can be used to calculate continuous variations in former ELAs as schematically illustrated in Fig. 6. For small glaciers only the most sensitive proglacial lake closest to the glacier can record fluctuations in GS/ELA, while for large glaciers only the most distant proglacial lake is sensitive for variations in GS/ELA (see Fig. 2).

If the investigated chain of proglacial lakes has a GS/SSD ratio which allows all variations in GS within a given time span to be recorded, error bars for the estimated ELAs of less than  $\pm 50$  m are suggested. Reconstructed ELAs must normally be adjusted for glacio-isostatic land uplift (e.g. Dahl and Nesje, 1996).

#### 4.3. One-site approaches

Downstream of many glaciers, suitable proglacial sites are often lacking or scarce. Hence, in many cases a setting with only one proglacial lake is all that is available to investigate how such glaciers and the corresponding ELAs have fluctuated backwards in time. If a glacier exists in the catchment at present, various proxies can record whenever former glaciers existed by using 'the modern analogue principle'. With a multi-site approach these proxies can be calibrated against independent sites with different sensitivity to record variations in glacier magnitude/ELA. For one-site approaches, variations in glacier activity/ELAs depend on the interpretation and sensitivity of the available methods (Fig. 2). However, some of these methods may be sensitive to record variations in small glaciers, while



others can be used to record fluctuations in larger glaciers.

## 5. Discussion

Effective rates of glacial erosion varies from  $0.01 \text{ mm yr}^{-1}$  for polar glaciers and thin temperate plateau glaciers on crystalline bedrock, to  $0.1 \text{ mm yr}^{-1}$  for temperate valley glaciers on resistant crystalline bedrock in Norway, to  $1.0 \text{ mm yr}^{-1}$  for small temperate glaciers on various bedrock types in the Swiss Alps, and to  $10\text{--}100 \text{ mm yr}^{-1}$  for large and temperate valley glaciers in the tectonically active mountain ranges of southeast Alaska (Hallet et al., 1996, and references therein). Hence, the link between variations in GS and the corresponding ELA based on proglacial lakes must be established for each glacier.

The bedrock beneath a glacier can be regarded as 'constant', whereas both temperature regime and thickness may vary with the size of the glacier. The temperature regime of a glacier also depends on air temperature and winter precipitation, and shifts from polar or polythermal to temperate may have taken place at the Younger Dryas/Holocene transition or from temperate to polythermal after the Holocene climatic optimum. Variations in the turnover time of ice in temperate glaciers may also have some influence on rates of effective glacial erosion.

The annual sediment transport along glacier meltwater streams normally exceeds by several orders of magnitude nonglacial streams with similar water discharge in Norway (e.g. Roland and Haakensen, 1985). This transport occurs as rolling, sliding and saltation along the channel bed, or in suspension. Some grains descending during saltation may be temporarily buoyed by upward movement in turbulent flow, and this condition can be described as incipient suspension. The weight of fine particles in true suspension is entirely supported by the upward pulses of flow generated by eddies (e.g. Summerfield, 1996). It is particles deposited from true suspension which make the ideal basis for using proglacial sites to reconstruct variations in glacier extent/ELA. Depending on the site and the competence of the meltwater stream, however, particles from incipient suspension are commonly found as coarser grains (coarse silt to sand) in sediments deposited at proglacial sites. For most proglacial sites, however, the glacier signal is found in fine- to medium silt (e.g. Leemann and Niessen, 1994; Matthews et al., 2000; Nesje et al., 2001). Hence, proglacial sites dominated by sedimentation from true suspension are preferred (Fig. 4).

In a study on the relationship between glacial activity and sediment production in the varved Hector Lake, Alberta in Canada, Leonard (1997) found that longterm

variations (century to millennial duration) in sedimentation rate reflected changes in glacier extent on the same timescale. However, decadal-scale variability more complexly related to upstream ice extent is superimposed on the longterm changes. High sedimentation rates were associated with glacier maximum positions, or with transitional periods preceding or post-dating periods of maximum ice extent. The glacier-covered area in the catchment of Hector Lake has varied from 60% during the Little Ice Age to 40% at present, a coverage of glaciers four to six times higher than for the majority of similar investigations in southern Norway.

The first lake in a chain of proglacial lakes acts as a sediment trap for coarser sediments. If the first lake is covered by the glacier, this is reflected as a shift from a low-energy mode to a high-energy mode in the lacustrine sedimentation of the second proglacial lake (e.g. Nesje et al., 2001). As temporary ice-dammed lakes commonly occur along glacier margins both during advance and retreat, this may explain some of the difficulties in interpreting whether high sediment production is directly linked to glacier maximum positions or not.

In a multidisciplinary study, Snowball and Sandgren (1996, 1997) recored proglacial lakes in the Kårsa valley in northern Sweden first investigated by Karlén (1976, 1981, 1997).

Based on different methodological approaches, Snowball and Sandgren (1996) found that following the last deglaciation, glaciers had existed in the catchment only for the last 3000  $^{14}\text{C}$  yr BP, a result which strongly contrasted the interpretation of Karlén (1976, 1981) who had suggested several glacier advances throughout the entire Holocene. Based on this investigation, Snowball and Sandgren (1996) strongly argued against single-core (site) studies. They also argued that only features that are consistently reproducible and can be dated in spatially distributed cores should be interpreted in terms of glacier activity, environmental conditions and climate change. However, the problems of getting reproducible results can normally be solved by using proglacial lakes with a flat bottom.

Brauer et al. (2001) compared four sediment profiles from lakes Holzmaar and Meerfelder Maar in the Eifel region, Germany. Based on varve-dating and pollen profiles from the two lakes, former discrepancies between the two lakes were explained after detailed correlation. They concluded that even in small lakes like Holzmaar discrepancies of several hundred years may occur, and that a multi-core study on two lakes from the same region is necessary to detect errors in single-core studies on nonvarved sediments.

Multi-core/site approaches are therefore preferred (e.g. Snowball and Sandgren, 1996; Brauer et al., 2001). However, suitable proglacial sites are difficult to find in many regions, and in many cases none or very

few sites are available. To minimize within lake variance and maximize between lake variance, basins distal from the inlet and/or the deepest part of the lake appear to give the best reproducible results when more than one core/site are taken into account. For one-site studies, only features that are consistently reproducible based on several independent proxies in two or more cores should be interpreted in terms of glacier variations and climate change. The principal coupling between various techniques/proxies and former glacier activity/ELAs is demonstrated in Fig. 2. Hence, for one-site approaches it is especially important that these principles are followed.

Church and Ryder (1972) defined the term ‘paraglacial’ as referring to “nonglacial processes that are directly conditioned by glaciation”. The term has been widely used to describe the reworking of glacial sediments by colluvial processes and running water after the withdrawal of glacier ice, including the landforms and sediment accumulations produced by such processes (e.g. Ballantyne, 1995). Attributed to nonglacial activity, thin ( $\leq 2$  cm) minerogenic layers and less-regular layers composed of coarse, angular sand and gravel particles, are found in both glacial and nonglacial lakes with steep slopes in the catchment (Matthews and Karlén, 1992). The thin layers are suggested to result from precipitation-induced events, including debris flows (Østrem and Olsen, 1987; Jonasson, 1991), while the less-regular layers are interpreted as ice-rafted colluvial debris (e.g. Luckman, 1975).

Colluvial activity often occurs within a limited area and as short (hours to days) events. However, similar minerogenic layers in nonglacial control lakes (Matthews and Karlén, 1992) and multi-core/site approaches reflecting the same glacier may reveal the origin of a nonglacial layer. Due to the longevity of many glacier-induced events (several hundred years) compared to colluvial events, radiocarbon dates above and below the actual layer may in some cases disclose the depositional agent.

By using ‘ward sorting’ on grain-size distributions to establish cumulative platforms, ‘true’ glacial meltwater sediments may be separated from deposits originating from colluvial activity (Blott and Pye, 2001).

## 6. Summary and conclusions

Except for historical records and observed mass-balance records, knowledge of former variations in glacier activity/ELAs rely, directly or indirectly, on the maximum altitude of lateral moraines and on information from proglacial lacustrine and terrestrial sites. As lateral moraines only reflect shorter periods when the glaciers obtained steady state in advanced positions beyond later glacier advances, continuous Holocene variations in glacier activity/ELAs can only be obtained

from proglacial sites beyond the Little Ice Age maximum.

In this paper, various approaches and techniques for reconstructing variations in former glacier activity/ELAs based on proglacial sites are evaluated, and criteria for site selection are discussed. The following conclusions and implications of systematic importance are proposed:

- (1) Records of glacier activity/ELAs obtained from proglacial sites are based on a conceptual model of glacier-meltwater-induced sedimentation in which the minerogenic (nonorganic) component of the sediments is related to the occurrence of a glacier in the catchment (Fig. 1) (e.g. Karlén, 1981; Leonard, 1985).
- (2) The principal coupling between various approaches and former glacier activity/ELAs is the same, and both possibilities and limitations are exemplified in Fig. 2. Problems in the interpretation and calibration of these parameters are primarily related to when the glacier was very small/melted away, or when it was so large that the used proxy lacked sensitivity to record variations in GS. However, the various approaches may complement each other.
- (3) Within the studied time span (e.g. the Holocene), the glaciated area in the catchment must be appropriate for recording the amplitude of variations in GS/ELA. The largest glacier in the catchment must also be classified (cirque glacier, plateau glacier, etc.) to understand better which climatic factors influence the local glacier activity/ELA (Fig. 3) (Dahl and Nesje, 1992; Dahl et al., 1997). Reconstructed ELAs must normally be adjusted for glacio-isostatic land uplift.
- (4) Catchments with a high number of proglacial lakes and other sites/features (local watersheds, lateral moraines, etc.) suitable to record variations in GS/ELAs are to be preferred (Figs. 4 and 5). Proglacial lakes should be dammed by a rock sill, and the shape of the lake basins should minimize post-depositional disturbance of the sediments. With no glaciers in the catchment, ‘proglacial’ lakes should have high organic production to increase the contrast, and the residence time of water in the proglacial lakes must allow both settling and further downstream transport of suspended sediments. Representative nonglacial control lakes should exist in the catchment (Matthews and Karlén, 1992), and geomorphological processes (colluvial activity, floods, etc.) which may influence on lake sedimentation must be taken into account.
- (5) Ideal proglacial sites turn out to record simple systems where the topographic conditions isolate the glacier meltwater signal through natural

filtering in a way that directly reflects GS/ELA. The occurrence of representative ice-marginal moraines of known age in the catchment is crucial for the calibration of glacier activity/ELAs based on proglacial sites.

- (6) Combined with proglacial sites, rerouting of glacier meltwater across local watersheds may give detailed information concerning former glacier activity/ELAs. Due to the sharp on-off signal, reconstructed variations in GS/ELA related to local watersheds can be used to calibrate ELAs based on a chain of proglacial sites (Fig. 4) (e.g. Dahl and Nesje, 1994; Dahl et al., 2002).
- (7) A close relationship between GS and downstream transport distance of glacier-induced sedimentation from true suspension is suggested based on Roland and Haakensen (1985). As a consequence, small glaciers can only be recorded at the most sensitive sites, while larger glaciers in addition can be recorded further downstream. If this catchment specific GS/SSD ratio is known (Fig. 6), records of glacier-induced sediments from a chain of proglacial lakes may give continuous sensitive variations in glacier magnitude backwards in time. GS is converted to an ELA estimate using the AAR-method. Independent observations (lateral moraines, local watersheds, etc.) are used to calibrate the amplitude of ELA fluctuations for catchment dependent factors. If the investigated chain of proglacial lakes has a GS/SSD ratio which allows all variations in GS within a given time span to be recorded, error bars for the estimated ELAs may be less than  $\pm 50$  m.
- (8) With a one-site approach, variations in glacier activity/ELAs depend on the interpretation and sensitivity of the used methods. Hence, to minimize within lake variance and maximize between lake variance, basins distal to the inlet and/or the deepest part of the lake appear to give the best reproducible results when more than one core is taken into account. For one-site studies, only features that are consistently reproducible based on several independent proxies in two or more cores should be interpreted in terms of glacier variations and climate change (e.g. Snowball and Sandgren, 1996; Brauer et al., 2001).
- (9) A critical factor for the use of both one-site approaches and a chain of proglacial lakes is the link between glacier advances and sediment production. Whether a longer transport length of sediments in suspension can be related to glacier maximum positions, or to periods preceding or post-dating periods of maximum ice extent (e.g. Leonard, 1997), is important for the interpretation of all proglacial sites, and must be further tested.

- (10) Comparison of reconstructions using approaches based on both a single proglacial lake and a chain of proglacial lakes for the same glacier is important for the development of reliable methods/techniques to reconstruct former glacier activity/ELAs. Hence, testing and improvement of relevant field and laboratory approaches must continue, and especially how various methods/techniques complement each other must be better understood.

### Acknowledgements

We thank John A. Matthews and an anonymous reviewer for comments on the manuscript. This paper is a contribution to NORPEC, a NFR funded Strategic University Program (SUP) at the University of Bergen, and to the Bjerknes Collaboration for Climate Research.

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## Paper 2

**Bakke, J., Dahl, S. O. and Nesje, A.** (submitted). Lateglacial and early-Holocene palaeoclimatic implications based on reconstructed glacier fluctuations and equilibrium-line altitudes at northern Folgefonna, Hardanger, western Norway. *Journal of Quaternary Science*.





# Lateglacial and early-Holocene palaeoclimatic implications based on reconstructed glacier fluctuations and equilibrium-line altitudes at northern Folgefonna, Hardanger, western Norway

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Submitted to *Journal of Quaternary Science*

**Abstract:** Northern Folgefonna (c. 23 km<sup>2</sup>), is a nearly circular maritime ice cap located on the Folgefonna Peninsula in Hardanger, western Norway. By combining marginal moraines and AMS radiocarbon dated glacier-meltwater induced sediments in proglacial lakes draining northern Folgefonna, a continuous high-resolution record of variations in glacier size and equilibrium-line altitudes (ELAs) during the Lateglacial and early-Holocene has been obtained. After the termination of the Younger Dryas (c. 11,500 cal. BP), a short-lived (150-200 years) climatically induced glacier readvance termed 'Jondal Event 1' occurred within the 'Preboreal Oscillation' (PBO) c. 11,200 cal. BP. Bracketed to 10,550-10,350 cal. BP, a second glacier readvance is named the 'Jondal Event 2'. A third readvance occurred about 10,000 cal. BP and corresponds to the 'Erdalen Event 1' recorded at Jostedalbreen. An exponential relationship between mean solid winter precipitation and ablation-season temperature at the ELA of Norwegian glaciers makes it possible to reconstruct former variations in winter precipitation if the corresponding ELA is known and combined with an independent proxy for summer temperature. Compared to at present, the Younger Dryas was much colder and drier, the 'Jondal Event 1'/PBO was colder and somewhat drier, the 'Jondal Event 2' was much wetter, whereas the 'Erdalen Event 1' started as rather dry and terminated as somewhat wetter. Variations in glacier magnitude/ELAs and corresponding palaeoclimatic reconstructions at northern Folgefonna suggest that low-altitude cirque glaciers (lowest altitude of marginal moraines 290 m asl.) in the area existed for the last time during the Younger Dryas. These low-altitude cirque glaciers of suggested Younger Dryas age do not fit into the previous reconstructions of the Younger Dryas ice sheet in Hordaland.

**Key words:** Younger Dryas, Preboreal Oscillation (PBO), Jondal Event, Erdalen Event, glacier fluctuations, equilibrium-line altitudes (ELAs), winter precipitation, grain-size analyses, Norway.

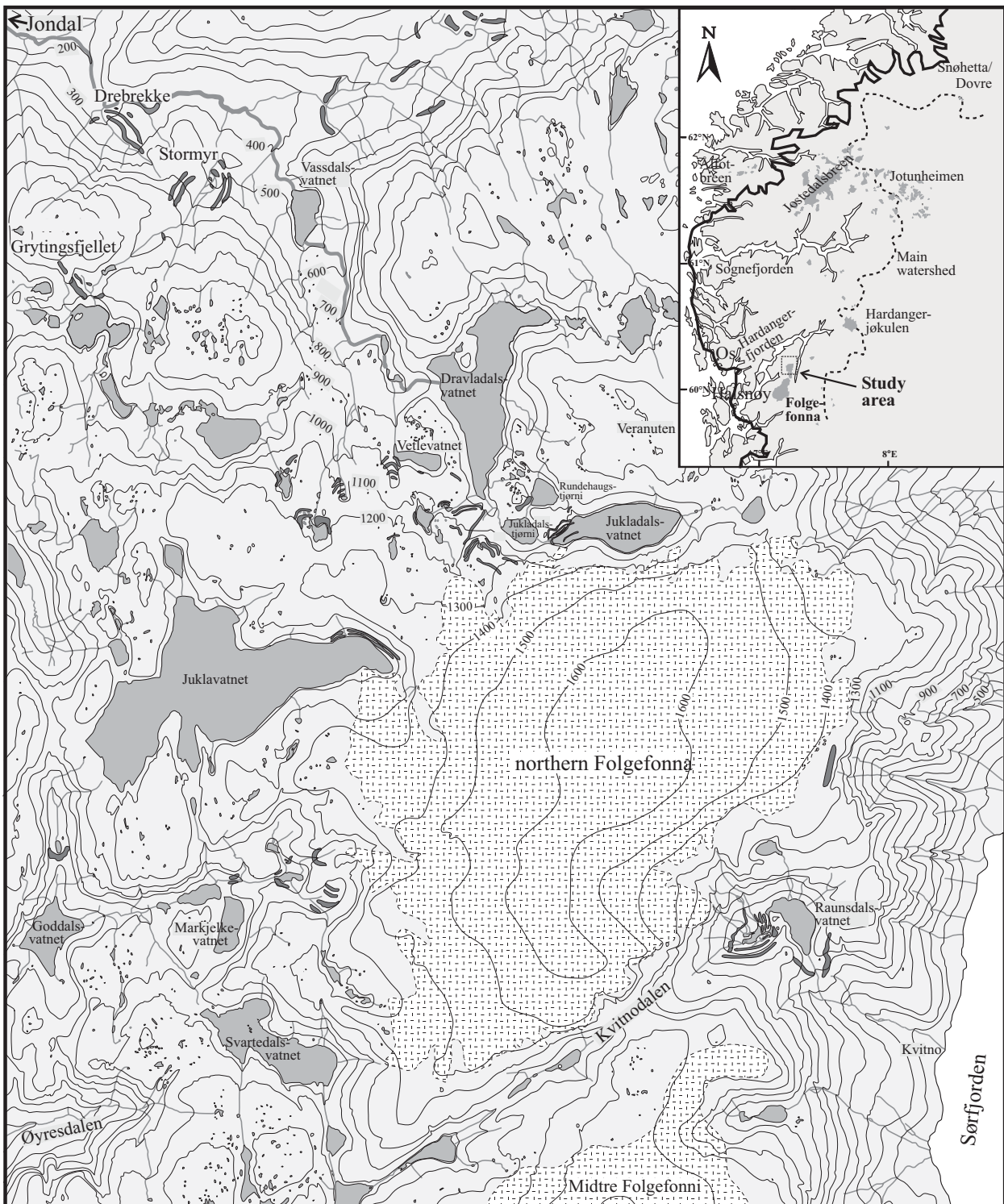
## Introduction

The transition from glacial to interglacial conditions is well understood. Of special interest is the general climatic implications of the Younger Dryas which close to 11,500 cal. yr BP terminates the last glacial stage and precedes the present interglacial, the Holocene.

A large number of Lateglacial and early-Holocene climate reconstructions are available from NW Europe. In western Norway climate reconstructions from the last deglaciation and early-Holocene have been inferred from records of Younger Dryas and early-Holocene glacier variations (e.g. Larsen *et al.*, 1984; Nesje *et al.*, 1991; Dahl and Nesje, 1992; 1994; 1996b; 2000a; 2001a; Dahl *et al.*, 2003), from biological proxies (e.g. Paus, 1988; 1989; Birks *et al.*, 1994; Birks and Ammann, 2000; 2000) and from variations in

weight loss-on-ignition (Nesje and Dahl, 2001). Off western Norway, marine climate records from the deglaciation are available from the southeast Norwegian Sea based on diatom data (Koc Karpuz and Jansen, 1992), and from the Troll area in the North Sea based on percentage variations of a cold water planktonic foraminifera (Klitgaard-Kristensen *et al.*, 2001).

The INTIMATE group suggested that the Greenland ice core GRIP should constitute the stratotype for the last termination (Björck *et al.*, 1998; Lowe *et al.*, 2001). These data have been used to calibrate different proxies from all over the Northern Hemisphere, and are also reproduced from European tree-ring chronologies (Friedrich *et al.*, 2001). A recent chironomid-based summer-temperature curve from Whitrig Bog, SE Scotland (Brooks and Birks, 2000) is in addition relevant for this study.



**Figure 1** The incised map shows the geographical distribution of glaciers in southern Norway. The dotted frame shows the study area. The main map is showing the ice cap northern Folgefonna and the surrounding area. Note the position of the proglacial lakes Vetlavtn and Vassdalsvatn, and the sites with marginal moraines deposited by low-altitude cirque glaciation at Drebrekke and Stormyr. Marginal moraines deposited by northern Folgefonna or by local cirque glaciers are marked as dark shaded lines. In the lower valley of Jondal there are some remnants of marginal moraines deposited by the Late Weichselian Scandinavian Ice Sheet.

Terrestrial archives reflecting the Lateglacial/early-Holocene transition in western Norway must be beyond the Younger Dryas continental ice sheet in western Norway (Fig. 1). Due to calving, the continental ice sheet is suggested to have retreated far inland along Hardangerfjorden prior to a major readvance in the order of 100 km during the Younger Dryas (Aarseth and Mangerud, 1974; Mangerud, 2000) (Fig. 1). However, reconstructed sea-level

fluctuations in inner parts of Hardangerfjorden indicate a final deglaciation prior to the Younger Dryas (Helle *et al.*, 1997). To have a major Younger Dryas readvance, the regional equilibrium-line altitude (ELA) must have been well below all high-lying mountain areas surrounding the wide and very deep Hardangerfjord. Of especial interest in this context is the Folgefonna Peninsula with three large plateau glaciers at present. If a readvance took place





**Figure 2** Photo showing the northern Folgefonna ice cap seen from southern Folgefonna to the south. The ice cap is almost circular and is located on a mountain plateau at an altitude of about 1300 m.

in Hardangerfjorden during the Younger Dryas, the Folgefonna Peninsula must have produced a large part of the glacier ice necessary to fill up the fjord. Findings of marginal moraines after low-altitude cirque glaciers down to 290 m asl. in Jondal well inside the suggested front of the Younger Dryas ice sheet in Hardangerfjorden have challenged this model (Bakke *et al.*, 2000).

By coring proglacial lakes draining northern Folgefonna and combine periods with AMS radiocarbon dated input of glacier-induced sediments with marginal moraines, the main objective of this paper is to reconstruct variations in glacier magnitude/ELAs and winter precipitation during the Lateglacial and early-Holocene in this region. We also want to discuss local implications of these reconstructions with respect to the age of marginal moraines after former low-altitude cirque glaciers in the area, and to discuss implications with respect to the regional ice-sheet configuration in Hardanger during the Lateglacial and early-Holocene. Finally, we want to establish an event chronology for the Lateglacial/early-Holocene transition based on glacier fluctuations/ELAs in western Norway.

## Study area

The ice cap northern Folgefonna covers an area of 23 km<sup>2</sup>, and is the seventh largest glacier in Norway. It has a circular configuration with an altitudinal range from 1644 to 1200 m, and a modern ELA of about 1465 m. There are five major outlet glaciers from the ice cap; Jordalsbreen, Juklavassbreen, Botnabreen, Dettébrea and Jukladalsbreen (Figs. 1 & 2). About 12 km<sup>2</sup> of northern Folgefonna drains

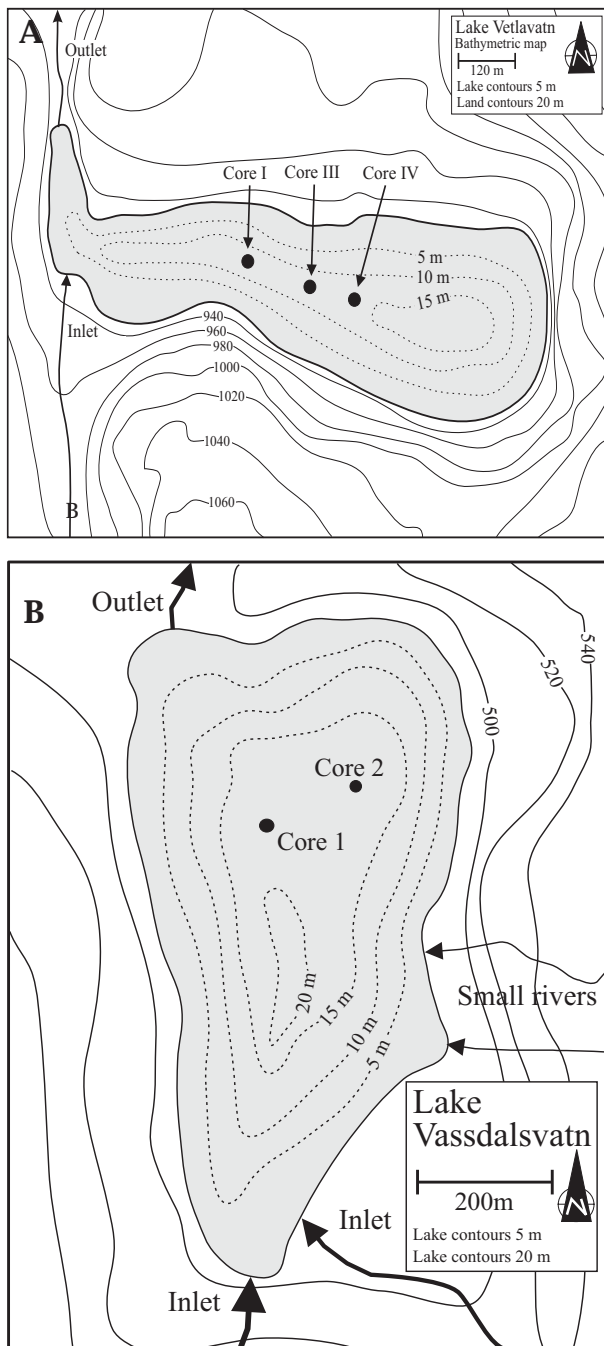
northwards, and this Jondal catchment includes 9 smaller and bigger lakes. No other glaciers exist in this catchment.

The bedrock in the upper Jondal catchment mainly consists of acid meta-andesite, metadacite, quartzite, migmatite and migmatitic schist of Precambrian age (Sigmond, 1985; Askvik, 1995). The vegetation is meagre around northern Folgefonna, and except for some marginal moraines in front of the major outlet glaciers, there is only a sparse cover of colluvium and till in the area (Bakke, 1999).

Based on a combination of two meteorological stations along Hardangerfjorden (Station no. 4949, Ullensvang Forsøksgård, 12 m a.s.l., 1962-1988; Station no. 5013, Omastrand, 1 m a.s.l., 1962-1990) (Klimaavdelingen, 1993b), the present mean summer temperature (Ts) from 1 May to 30 September is 12.7°C at sea level in Jondal. Using an environmental lapse rate of 0.6°C/100 m (e.g. Sutherland, 1984; Dahl and Nesje, 1992), this gives a mean Ts close to 4.0°C at the modern ELA of 1465 m at northern Folgefonna. At a local meteorological station about 11 km from the present glacier terminus of northern Folgefonna (Station no. 5696, Kvåle, 342 m a.s.l., 1961-1990) (Klimaavdelingen, 1993a), the present mean winter precipitation (Pw) from 1 October to 30 April is 1434 mm. Based on a mean observed exponential increase in winter precipitation with altitude of 8%/100 m in southern Norway (Haakonsen, 1989; Dahl and Nesje, 1992, and references therein), the corresponding Pw at the ELA of northern Folgefonna is c. 3350 mm.

## Research approach and methods

The reconstruction of Lateglacial and early-Holocene glacier fluctuations at northern Folgefonna involved



**Figure 3** (A) Bathymetric map of lake Vetlavatn showing the location of the retrieved piston cores. Core I to IV is taken with increasing distance away from the inlet/outlet. Northern Folgefonna did not reroute glacier meltwater to lake Vetlavatn during the 'Little Ice Age'. (B) Bathymetric map for lake Vassdalsvatn showing the location of the retrieved piston cores. Note that core I is taken close to the main 'river stream channel' running through the lake.

several approaches:

- Air photographs (Widerøe, 1962) and field observations were combined to produce a detailed glacial-geomorphological map in 1:5000 for the Jondal catchment with special emphasis on former marginal moraines, glacier-meltwater channels, glaciofluvial deposits, marine terraces and various ice-flow indicators.

- Primarily to sort out moraines older than the historical 'Little Ice Age' maximum, lichenometry

based on *Rhizocarpon geographicum* (Matthews, 1994) and Schmidt-hammer rebound values according to McCarroll (1994), were used to establish relative age chronologies for the marginal moraines in front of three outlet glaciers from northern Folgefonna and at moraines after two former low-altitude cirque glaciers.

- The accumulation of distinct marginal moraines is suggested to be closely related to (rather) short periods when a glacier was in steady state, and younger glacier advances may also erase older marginal moraines (e.g. Matthews, 1991; Dahl and Nesje, 1994). Hence, proglacial sites (lacustrine and terrestrial) beyond the maximum extent of the suggested glacier advances in the studied time span are taken into account to obtain continuous records of glacier fluctuations. Various methods related to proglacial sites are all based on a conceptual model of glacier-meltwater induced sedimentation in which the minerogenic (nonorganic) component of the sediments is related to the occurrence of a glacier in the catchment (Karlén, 1981; Leonard, 1985; Dahl *et al.*, 2003). In this study, the two proglacial lakes Vetlavatn and Vassdalsvatn (Figs. 1&3) were cored in an attempt to obtain absolute dating control on the timing and magnitude of Lateglacial and early-Holocene glacier fluctuations in the Jondal catchment.

Lake Vetlavatn, at an altitude of 915 m asl., covers an area of 0.1 km<sup>2</sup> (Fig. 3a), and is situated in a glacially eroded bedrock basin with the longest axis oriented east-west in a length of 700 m. The proglacial lake only receives glacier-meltwater induced sediments when the outlet glacier Jordalsbreen from northern Folgefonna advances beyond a local bedrock threshold. When the glacier is behind this local watershed, organic gyttja dominates the sedimentation in the lake. Four cores were retrieved from this lake.

Lake Vassdalsvatn, at an altitude of 490 m asl., covers an area of 0.17 km<sup>2</sup> (Fig. 3b), and is the seventh proglacial lake downstream from northern Folgefonna along the present meltwater stream in Jondal. The lake is located in a glacially eroded bedrock basin, and it has an input of glacier-meltwater induced sediments at present. Lake Vassdalsvatn is suggested to reflect for whenever northern Folgefonna is very small or melted completely away. Two cores were retrieved from lake Vassdalsvatn.

Both proglacial lakes were cored using a modified piston corer taking up to 6 m long cores with a diameter of 110 mm (Nesje, 1992). The laboratory analysis included weight loss-on-ignition (LOI) (Heiri *et al.*, 2001), bulk density (wet and dry) and water content (Menounos, 1997), and grain-size analysis using a Micromeritics Sedigraph 5100 (x-ray determination). See Dahl *et al.* (2003) for an evaluation of the principal use of these techniques to record former glacier fluctuations.

Seventeen and two bulk AMS radiocarbon dates



were carried out from lakes Vetlavatn and Vassdalsvatn, respectively. Terrestrial plant macrofossils for AMS radiocarbon dating were sparse or lacking at both sites. As both lakes are located in acid Precambrian granite gneiss, however, this is not regarded to be a problem for age-depth modelling (Barnekow *et al.*, 1998; Lowe and Walker, 2000). All results are shown in Table 1, and are presented in the text as calibrated years before present (cal. BP) according to INTCAL 98 (Stuiver *et al.*, 1998) if not otherwise stated.

The estimates of former glacier ELAs are based on observations of modern analogues and accumulated knowledge from previous works (Dahl *et al.*, 2003). Calculations at the plateau glacier are made by using an Accumulation Area Ratio (AAR) of 0.7 (Dahl and Nesje, 1996a), whereas the cirque glaciers are reconstructed using an AAR of 0.6 (Dahl *et al.*, 1997). The calculation of the area distribution was carried out electronically using the vector based GIS program MapInfo 6.0 on an N-50 map datum.

## Results

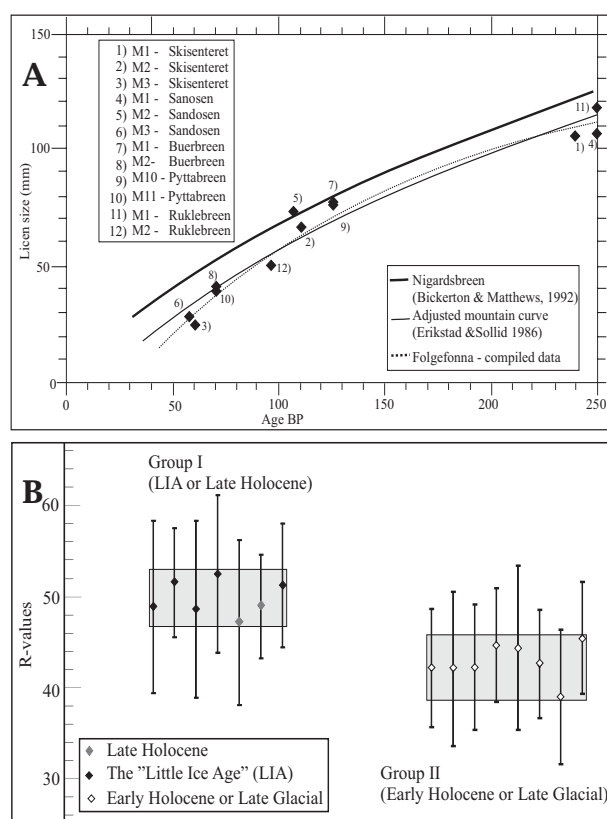
### Moraine chronology

The moraine chronology in front of outlet glaciers from northern Folgefonna indicates up to eight successively smaller glacier advances/readvances with deposition of marginal moraines. Beyond these moraines there are some sparsely distributed remnants after an even older glacier advance. Suggested to be the result of differences in aspect and slope, the moraine chronology is not consistent around northern Folgefonna.

A relative moraine chronology was established by use of lichenometry (Fig. 4a) and Schmidt-hammer rebound values (Fig. 4b) on the moraines. Based on a compilation of lichen measurements performed at the proximal side of marginal moraines around northern Folgefonna (Tvede, 1972; Bjelland, 1998; Bakke, 1999; Simonsen, 1999), three major glacier advances which occurred during the 'Little Ice Age' culminated about AD 1750, 1870 and 1930, respectively (Fig. 4a).

Based on observations in southern Norway, lichen-growth curves reflecting *Rhizocarpon geographicum* can not be used for reliable (relative) dating for more than 400-500 years backwards in time (Matthews, 1994). Hence, lichen diameters on the remaining five sets of marginal moraines frequently indicate an age of formation well beyond this time interval. Calibrated against the 'Little Ice Age' moraines, Schmidt-hammer rebound values (Fig. 4b) indicate that two marginal moraines may have formed c. 3,000-1,000 years ago, whereas the rebound values for the remaining three sets of marginal moraines suggested an age of deposition during the Lateglacial or early-Holocene.

Marginal moraines suggested to have been



**Figure 4** (A) Growth curves for the lichen *Rhizocarpon geographicum* in the marginal zones of Nigardsbreen and a composite mountain curve based on the curve from Middalsbreen at Hardangerjøkulen fit well with observations from different parts of the Folgefonna glacier. Based on the suggested growth rate, an early 'Little Ice Age' (LIA) glacial advance took place at Folgefonna close to 1750 AD, whereas marked later readvances occurred between AD 1870 - 1890 and during the AD 1930s. (B) Schmidt-hammer rebound values (R-values) from marginal moraines at the northern part of the ice cap northern Folgefonna. The figure shows how marginal moraines deposited during the Lateglacial and early-Holocene can be separated from moraines formed during the late-Holocene based on R-values.

formed by two low-altitude cirque glaciers have been investigated at Drebrekke and Stormyr in Jondal (Fig. 1). Below the mountain Grytingsfjellet (alt. 1092 m asl.), two sets of marginal moraines and indications of a third have been deposited by a small cirque glacier formed in a poorly developed cirque at Drebrekke. The eastern lateral moraine is well marked and can be traced to an altitude of 400 m, while the outer terminal moraine can be mapped to a lowest altitude of 290 m asl. (Figs. 5 & 6).

The farm inside the marginal moraines at Drebrekke has a history which can be followed back to the 15th century (Kolltveit, 1953). This suggests that no glacier formed at Drebrekke during the 'Little Ice Age', and this is confirmed by lichen measurements on the marginal moraines. Schmidt-hammer rebound values indicate an age of formation which is comparable with the three oldest sets of marginal moraines around northern Folgefonna.

At the northern flank of the mountain Storafjell (alt. 1133 m), two distinct sets of marginal moraines formed by a rather large cirque glacier are situated



**Figure 5** Photo showing the Drebrekke farm in Jondal. The Drebrekke farm lies on marginal moraines formed by a small cirque glacier of suggested Younger Dryas age N-NE of the mountain Grytingsfjellet (proximal side to the right on the photo). The lowermost part of the outer moraine ends at an altitude of 290 m. Hence, the Drebrekke site gives the maximum elevation for a fjord glacier if there existed a Hardangerfjord glacier during the Younger Dryas in this region (see Fig. 14).

north of a mature cirque at Stormyr (Fig. 1). Taken into account 4-5 m of peat, the moraines have a height of 7-11 m. The northwestern lateral moraine can be traced to an altitude of 590 m. Lichen measurements and Schmidt-hammer rebound values on the marginal moraines indicate a similar age of formation as at Drebrekke (Fig. 4b).

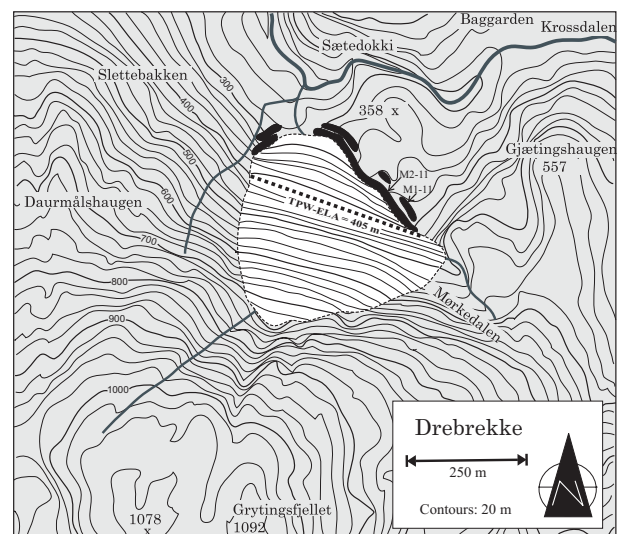
## Lithostratigraphy and radiocarbon dates

### Lake Vetlavatn

Based on cores I, III and IV from lake Vetlavatn, the combined lithostratigraphy may be subdivided into ten units, with core 4 as the most representative (Figs. 7 & 8). The content of cumulative medium silt including weight LOI from cores III and IV are shown in Fig. 9. Minor lithological variations between the cores are suggested to primarily reflect increasing distance away from the stream inlet/outlet in the western part of the lake (Fig. 3a). Details concerning the AMS radiocarbon dates used for age-depth control are shown in Table 1.

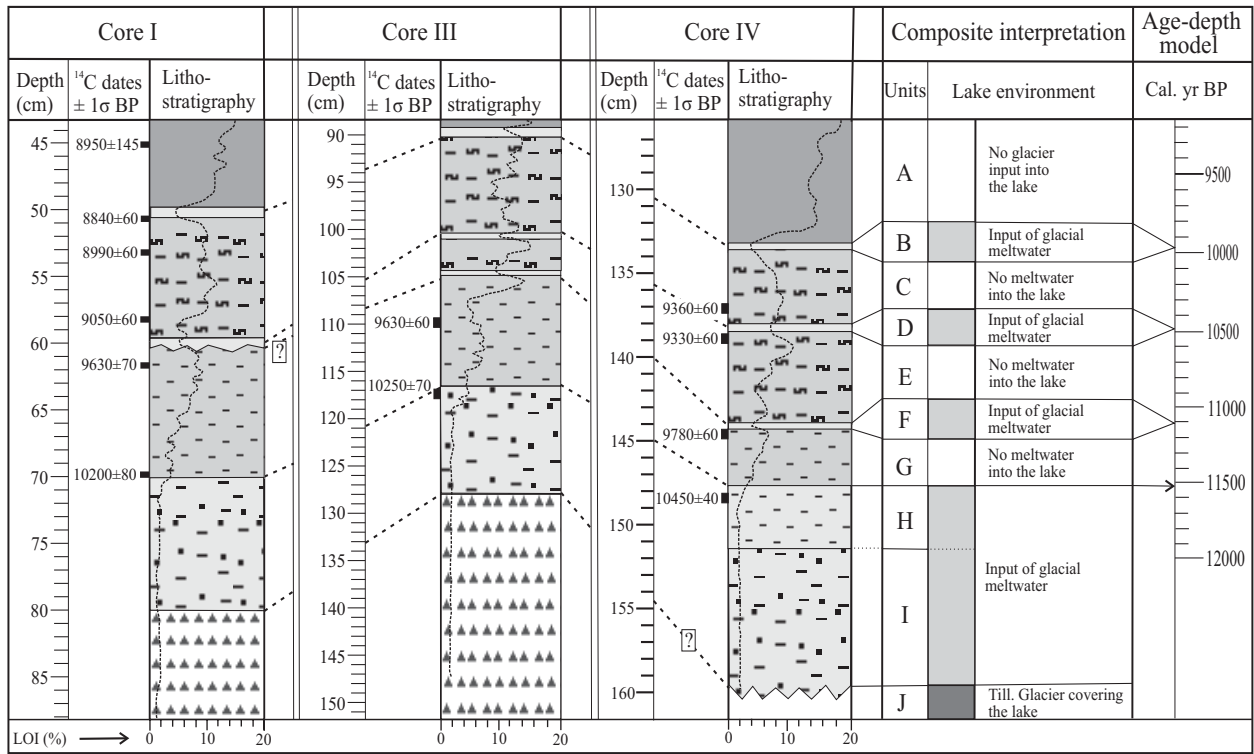
Unit J consists of a grey diamicton with angular gravel-sized particles in a matrix of silt and clay, and it has an average LOI of 0.8%. In core I, the unit is running from 87.5 to 80 cm, in core III from 151 to 129 cm, while it is missing in core IV. Unit I consists

of a light-grey coarse silt to sand, turning into the light-grey Unit H consisting of silt. The LOI varies from a minimum of 0.8% in Unit I to a maximum in Unit H of 4.5%. The end of Unit H is dated to



**Figure 6** The reconstructed cirque glacier at Drebrekke in Jondal with a temperature-precipitation-wind equilibrium-line altitude (TPW-ELA) of 405 m, which is a lowering of the ELA of 1150 m compared with the present situation at the ice cap northern Folgefonna. The extension of the glacier is based on lateral moraines and cirque topography. The contours are reconstructed using the shape of modern cirque glaciers in Norway.

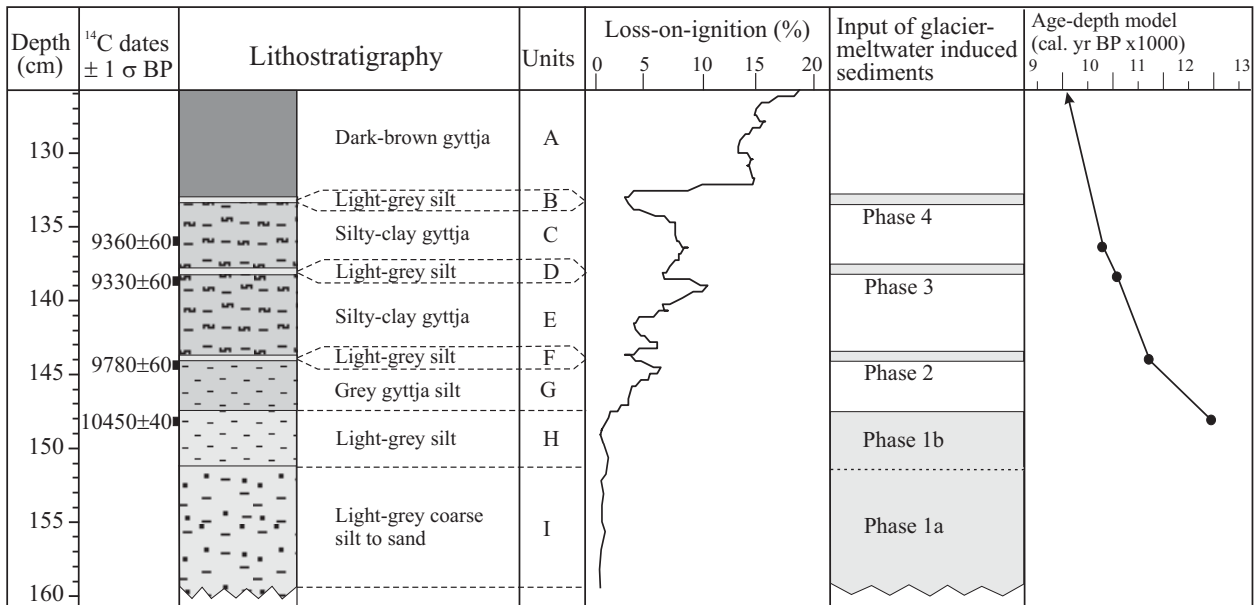




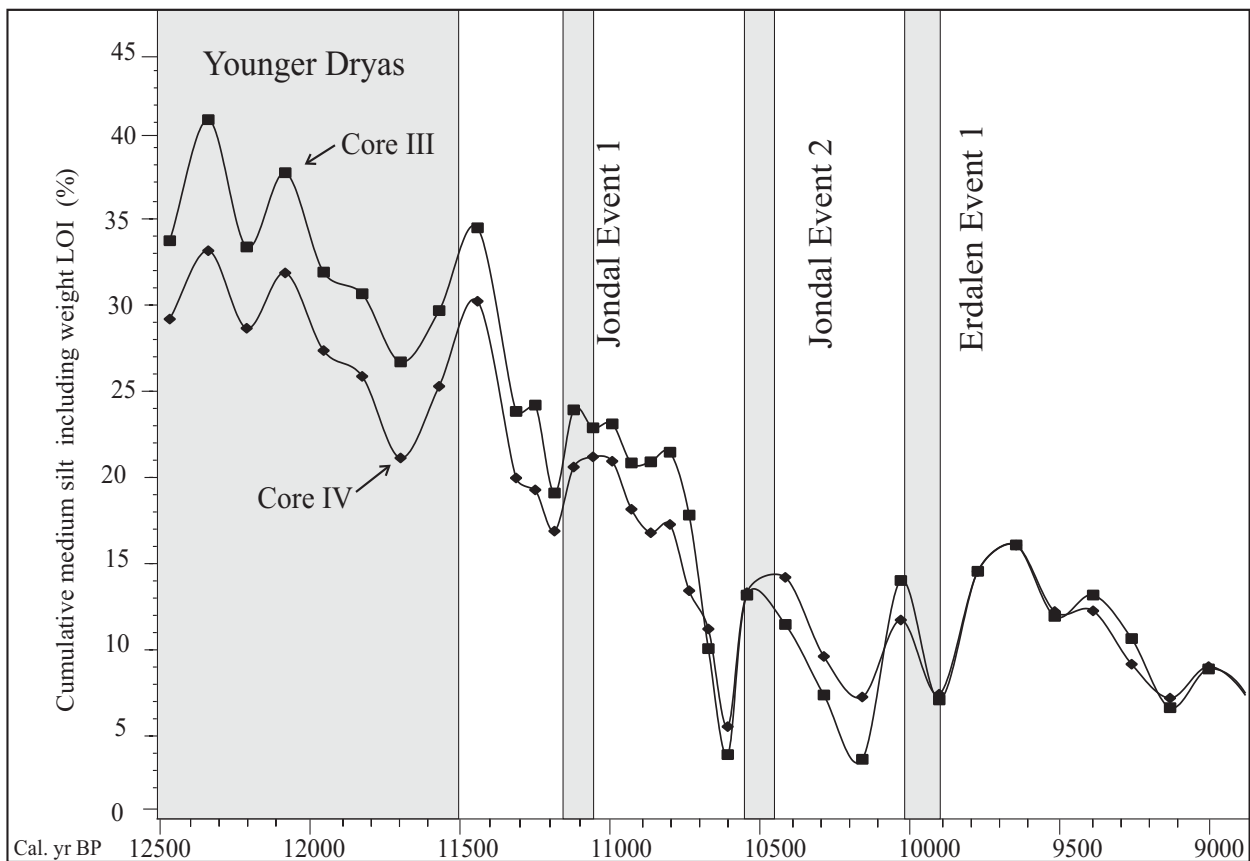
**Figure 7** Compiled lithostratigraphy from three cores derived from lake Vetlavatn. The radiocarbon dates are shown in radiocarbon years before present together with independent depth scales (cm) for each core. The dashed lines show the correlation between lithostratigraphical units and the right column the suggested age for the different units in calibrated years before present (=1950). Probably because of erosion by the river, the units H, G and F are missing in core I.

10,480±40 <sup>14</sup>C yr BP (Beta-148429) at 148 cm in core IV. The grey Unit G consists of gyttja silt with LOI values from 4.5 to 8% dated to 10,200±80 14C yr BP (Beta-115403) at 69.5 cm in core I and to 10,250±70 14C yr BP (Beta-148431) at 118 cm in core III. In Unit F light-grey silt turns into a more greyish colour upwards with a mean LOI of 4.0%. The 5 mm thick layer appears in sharp contrast to surrounding units, and it is present and radiocarbon dated at 109.5 cm

in core III to 9630±60 <sup>14</sup>C yr BP (Beta-148430), and at 144 cm in core IV to 9830±60 14C yr BP (Beta-148428), respectively. Unit E consists of grey silty-clayey gyttja that is turning into brownish yellow upwards, and with LOI values from 4.0 to 12.4%. A radiocarbon date from this unit yielded an age of 9660±70 <sup>14</sup>C yr BP (Beta-115403) at 61.5 cm in core I. In Unit D, a distinct 5 mm thick layer of light-grey silt has a mean LOI of 3.7%, and is bracketed in core IV by



**Figure 8** Lithostratigraphy with radiocarbon dates and environmental interpretation from Core IV in lake Vetlavatn. Note the changing sedimentation rate upwards in the core.



**Figure 9** Cumulative medium silt against loss-on-ignition for core III and IV in lake Vetlavtn on a calendar year scale. The figure shows that the input of silt decreased after the Younger Dryas with small peaks during the early-Holocene events. Probably because of increased precipitation, the content of medium silt increases after the 'Erdalen Event 1'.

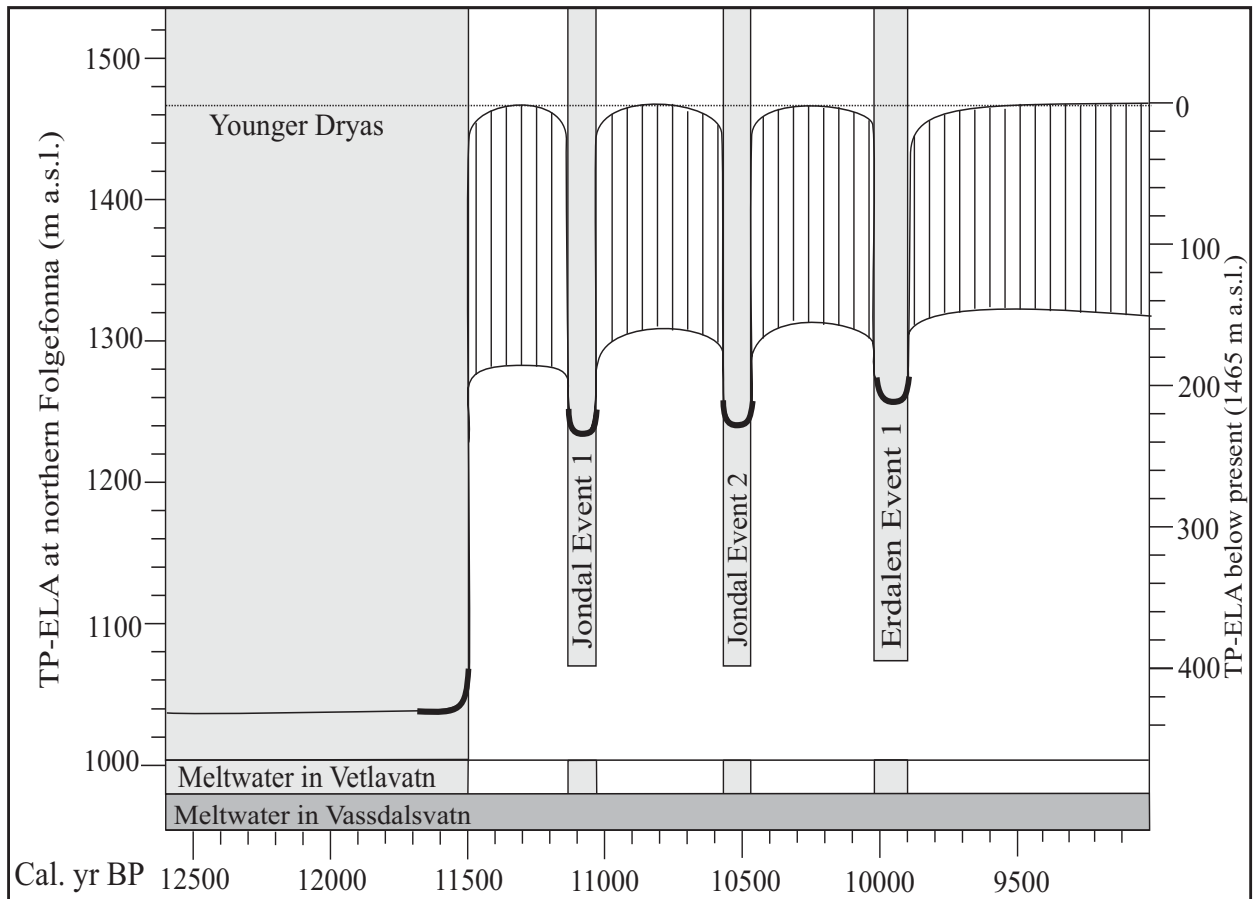
radiocarbon dates of  $9380 \pm 60$   $^{14}\text{C}$  yr BP (Beta-148427) and  $9360 \pm 60$   $^{14}\text{C}$  yr BP (Beta-148426) at 138 and 136 cm, respectively. Unit C consists of grey silty-clay gyttja with LOI values varying from 7 to 10%, whereas Unit B is a distinct 5 mm thick layer of light-grey silt with a mean LOI of about 3.0%. The unit is bracketed by radiocarbon dates in core I of  $9050 \pm 60$

$^{14}\text{C}$  yr BP (Beta-115401) at 58 cm, of  $8990 \pm 60$   $^{14}\text{C}$  yr BP (Beta-115401) at 53 cm and by  $8840 \pm 60$   $^{14}\text{C}$  yr BP (Beta-115401) at 50 cm. The homogeneous brownish gyttja representing Unit A is found in all three cores, and it has LOI values varying from a maximum of 31 % to a minimum of about 6.4%. It is radiocarbon dated in core I to  $7640 \pm 135$   $^{14}\text{C}$  BP (T-13605) at 33 cm, to

**Table 1** Radiocarbon dates from Vetlavtn. All dates are AMS bulk dates.

Core	Depth	Laboratory number	Radiocarbon age	Intercept	2 sigma cal. yr BP	$\delta^{13}\text{C}$
Vetlavtn I	15	T-13603A	$6785 \pm 160$	7680	7934 - 7468	-29.6
Vetlavtn I	20	T-13604A	$7475 \pm 26$	8482	8783 - 7789	-29.9
Vetlavtn I	33	T-13605	$7640 \pm 135$	8402	8502-8315	-27.9
Vetlavtn I	46	T-13606	$8950 \pm 145$	9959	10034-9859	-29.1
Vetlavtn I	50	Beta-115399	$8840 \pm 60$	9890	9920-9850	-27.5
Vetlavtn I	53	Beta-115400	$8990 \pm 60$	9975	10005-9940	29.0
Vetlavtn I	58	Beta-115401	$9050 \pm 60$	10005	10035-9975	-28.4
Vetlavtn I	61.5	Beta-115403	$9660 \pm 70$	10920	10960-10625	-28.8
Vetlavtn I	69.5	Beta-115403	$10200 \pm 80$	11960	12155-11680	-27.6
Vetlavtn III	110	Beta-148430	$9630 \pm 60$	8890	11160-10690	-28.1
Vetlavtn III	118	Beta-148431	$10250 \pm 70$	11940	12360-11580	-26.9
Vetlavtn IV	23	Beta-148424	$2980 \pm 40$	3160	3260-3000	-25.2
Vetlavtn IV	118	Beta-148425	$8150 \pm 50$	9020	9130-8990	-28.1
Vetlavtn IV	136	Beta-148426	$9360 \pm 60$	10520	10670-10270	-28.3
Vetlavtn IV	138	Beta-148427	$9380 \pm 60$	10540	10690-10380	-28.3
Vetlavtn IV	144	Beta-148428	$9830 \pm 60$	11190	11250-11130	-28.1
Vetlavtn IV	148	Beta-148429	$10480 \pm 40$	12500	12820-12080	-26.6
Vassdalsvatn II	171	UtC-6695	$6280 \pm 60$	7195	7220-7095	-27.1
Vassdalsvatn I	368-372	Beta-102935	$8260 \pm 80$	9235	9375-9025	-25.0





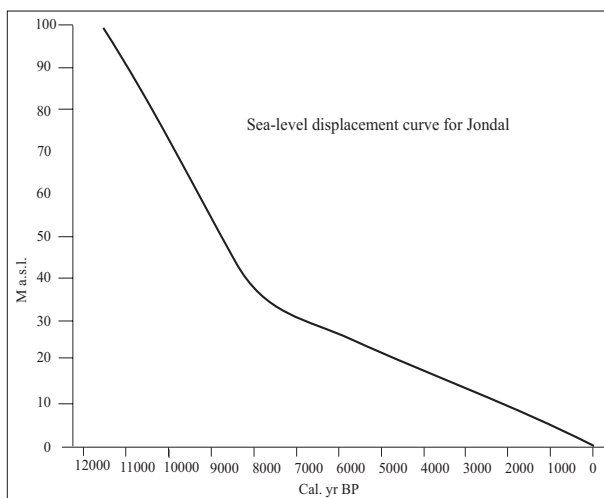
**Figure 11** Reconstructed Lateglacial and early-Holocene glacier fluctuations and TP-ELAs at northern Folgefonna. Note that the TP-ELA curve is adjusted for land uplift. Vertical bars indicate possible ELA altitude range during periods without meltwater input to lake Vetlavatn.

7475±30 <sup>14</sup>C yr BP (T-13604A) at 20 cm and to 6785±160 <sup>14</sup>C yr BP (T-13603A) at 15cm. A sample taken from Unit A at 46 cm in core I was radiocarbon dated to 8950±145 <sup>14</sup>C yr BP (T-13606), and is suggested to be somewhat too old. However, the deviation may be explained by a large standard

deviation. In core IV, Unit A is dated at 8150±50 <sup>14</sup>C yr BP (Beta-148425) at 118 cm and to 2980±40 <sup>14</sup>C yr BP (Beta-148424) at 23 cm.

**Lake Vassdalsvatn**

For this study, the lithostratigraphy in Lake Vassdalsvatn is primarily used to complement lake Vetlavatn and to indicate whether northern Folgefonna was melted away during the studied time span or not. Light-grey clayey silt is suggested to reflect the existence of northern Folgefonna in the catchment, whereas gyttja is likely to indicate whenever the glacier was melted away. The first transition from light-grey glacier-meltwater induced clayey silt to gyttja after the deglaciation is dated at 7200 cal. yr BP (UtC-6695) in core II, whereas a sediment layer rich in plant macrofossils is interpreted to be a flood deposit dated to 9200 cal. yr BP (Beta-102935) in core I (see Table 1 for details). The suggested flood eroded upstream peat deposits with resedimentation in the central basin/channel of lake Vassdalsvatn where core I is retrieved (see Bakke *et al.* in prep. for further discussion).



**Figure 10** Sea-level displacement curve for Jondal based on the equidistant sea-level (ESL) diagram from Hamborg (1983) used for adjusting the winter precipitation estimates during the early-Holocene at northern Folgefonna.

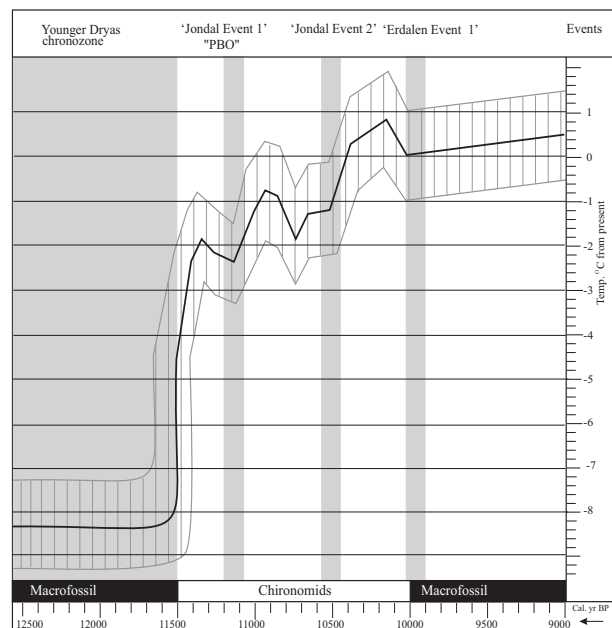
## Glacier variations and equilibrium-line altitudes at northern Folgefonna

The Lateglacial and early-Holocene glacier fluctuations are primarily based on a combination of the radiocarbon dated lithostratigraphies from lakes Vetlavatn and Vassdalsvatn and the moraine chronology at northern Folgefonna. The record from lake Vassdalsvatn indicates whenever northern Folgefonna existed during the investigated time span, whereas lake Vetlavatn record indicates whenever the glacier advanced beyond a local watershed which prevents direct input of glacier-meltwater induced sediments from northern Folgefonna to this lake at present. Hence, by combining these records with the moraine chronology it is possible to reconstruct the number, age and magnitude of Lateglacial and early-Holocene glacier fluctuations at northern Folgefonna.

The present temperature-precipitation-wind ELAs (TPW-ELAs) at the northern outlets of northern Folgefonna is 1465m, whereas estimates from the outlet glaciers to the west (Botnabreen) and to the east (Dettebrea) show TPW-ELAs of 1465 m and 1460 m, respectively. This demonstrates that the mean value of all local TPW-ELAs at northern Folgefonna may be regarded as a regional temperature-precipitation ELA (TP-ELA), which makes it possible to neglect the input of leeward accumulation of dry snow by wind (see Dahl and Nesje, 1996 and Dahl *et al.*, 1997; 2003, and references therein for definitions and further discussion).

By using an AAR of 0.7 for plateau glaciers in steady state (e.g. Dahl and Nesje, 1996a), it is possible to transform the reconstructed outlet glaciers into corresponding TP-ELAs. During the 'Little Ice Age' the TP-ELA is estimated to 1360 m in all three aspects. This gives a lowering of the TP-ELA of 105 m compared to the present value. For the Lateglacial and early-Holocene glacier events it is difficult to separate out the different ELAs based on the marginal moraines alone as these are located very close to each other. If not adjusted for glacio-isostatic land uplift, the estimated TP-ELA was close to 1220 m (lowering 245 m) during the suggested Younger Dryas and about 1320 m (lowering 145 m) during the early-Holocene glacier advances compared to at present. In general, the reconstructed glacier variations at northern Folgefonna are related to the local topography, while the reconstructed ELAs are adjusted for glacio-isostatic land uplift (Fig. 10) (Dahl and Nesje, *op. cit.*; Dahl *et al.*, 2003).

The former cirque glacier at Drebrekke has been reconstructed based on marginal moraines, and the former TPW-ELA has been calculated by the use of an AAR of 0.6 to 405 m (Fig. 6). Adjusted for a Holocene glacio-isostatic land uplift of 100 m (Fig. 10), the TPW-ELA lowering compared to the present TP-ELA of 1465 m at northern Folgefonna is 1160 m.



**Figure 12** Compiled ablation-season temperature curve for the Lateglacial and early-Holocene in western Norway based on Birks and Ammann (2000) and adjusted for land uplift.

The similar reconstructed TPW-ELA at the former Stormyr cirque glacier is 730 m. Adjusted for a Holocene glacio-isostatic land uplift of 100 m (Fig. 10), the lowering of the TPW-ELA compared to the present TP-ELA of 1465 m at northern Folgefonna is 835 m.

Lateglacial and early-Holocene variations in glacier magnitude at northern Folgefonna and the corresponding TP-ELA adjusted for glacio-isostatic land uplift are shown in Figure 11.

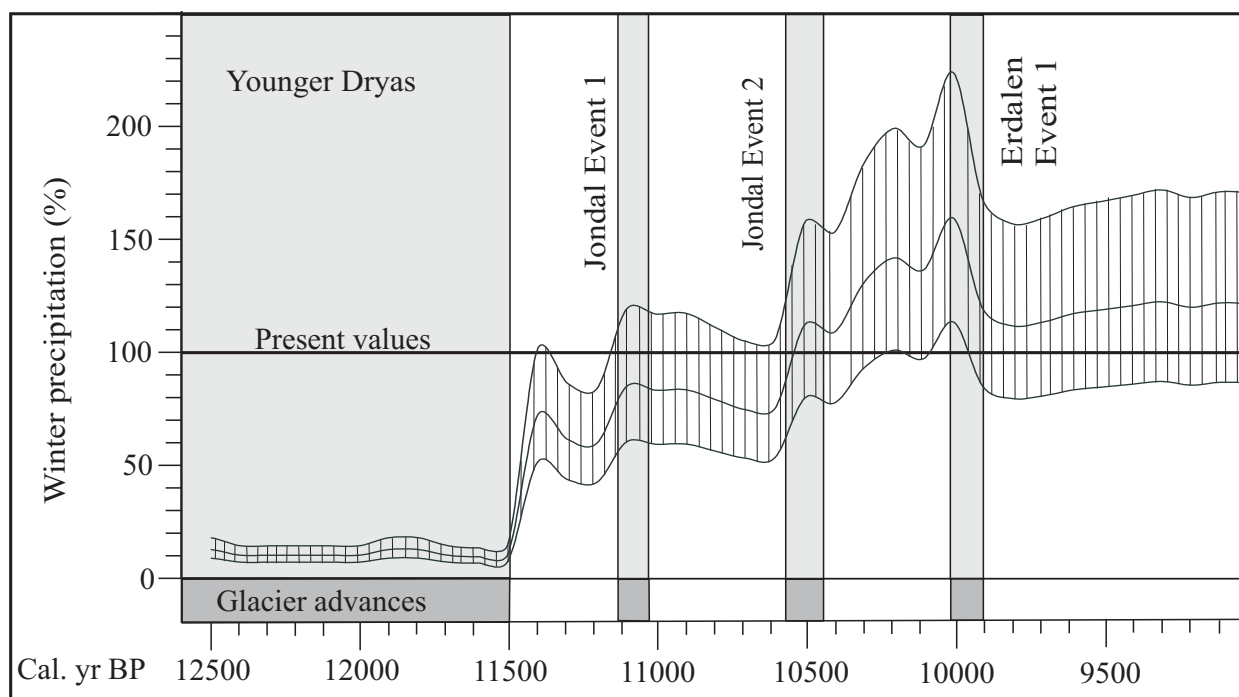
### Prior to the Younger Dryas (unit J)

The basal diamicton in Lake Vetlavatn is suggested to be a basal till deposited by a warm based glacier (Fig. 7). When the basal till was deposited is not known.

### Glacier magnitude and TP-ELA c. 12,500-11,500 cal. yr BP (the Younger Dryas)

Units I/H were probably deposited when the glacier was situated at the distinct terminal moraine at the western lake shore of Vetlavatn (Fig. 1). The sediments are characterized by a low organic content and a high proportion of coarse silt, and the radiocarbon dates are well within the middle to later part of the Younger Dryas (see Beta-148429, Beta-148131 and Beta-115403 in Table 1 and Fig. 7). A low amount of glacier meltwater derived sediments and low sedimentation rates are suggested to be the result of a cold based/polythermal glacier in the zone of continuous permafrost during this stage.

The TP-ELA lowering adjusted for land uplift is estimated to be 335 m during the suggested Younger Dryas.



**Figure 13** Holocene variations in winter precipitation derived from the 'Liestøl-equation' for the ice cap northern Folgefonna. The middle line is the average estimate based on the temperature data. Maximum and minimum estimates for the calculations are based on variations in the temperature-precipitation ELA (TP-ELA). The Vassdalsvatn core indicates that northern Folgefonna has existed during the entire investigated time span (TP-ELA lower than or equal to 1500 m), while Vetlavatn sorts out the events where the TP-ELA are close to 1320 m. For the rest of the time span the TP-ELA varies between 1500 and 1350 m.

#### **Glacier magnitude and TP-ELA c. 11,500-11,150 cal. yr BP**

Just prior to or when Unit G was deposited in lake Vetlavatn, the glacier is suggested to have retreated from the western lake shore to behind the local threshold further upstream. The high minerogenic content in Unit G is probably because of reworking of older glacier-derived sediments by fluvial erosion (Ballantyne, 1995). After an early maximum, the minerogenic input to the lake shows a gradual decrease during this period (Fig. 9). The TP-ELA lowering adjusted for land uplift is estimated to be 190 m during the time span 11,500-11,150 cal. yr BP.

#### **Glacier magnitude and TP-ELA c. 11,150-11,050 cal. yr BP (the 'Jondal Event 1'/PBO)**

Because of low LOI values, a relatively high content of coarse silt and high sedimentation rates, a glacier advance beyond the local watershed upstream of lake Vetlavatn is suggested to have occurred when Unit F was deposited. The glacier meltwater pulse (phase 2) representing Unit F is found in cores III and IV, while the lack of this unit in core I is suggested to be the result of later fluvial erosion close to the inlet (Fig. 3a). AMS radiocarbon dates also indicate that there is a hiatus in core I during this period. The TP-ELA lowering adjusted for land uplift is estimated to be 230 m during the 'Jondal Event 1'.

#### **Glacier magnitude and TP-ELA c. 11,050-10,550 cal. yr BP**

Just after the 'Jondal Event 1', the glacier retreated behind the local watershed upstream of lake Vetlavatn, and Unit E with a low content of coarse

silt and relatively high LOI values has no indications of glacier meltwater input to the lake. Suggested to be the result of later fluvial erosion near the inlet, Unit E is present in core IV, is found as a minor sequence in core III, and is lacking in core I (Figs. 3a & 7). The TP-ELA lowering adjusted for land uplift is estimated to be 160 m during the time span 11,050-10,550 cal. yr BP.

#### **Glacier magnitude and TP-ELA c. 10,550-10,450 cal. yr BP (the 'Jondal Event 2')**

Based on low LOI values, a relatively high content of coarse silt and increased sedimentation rates, the glacier had advanced across the local watershed and glacier meltwater was rerouted into lake Vetlavatn when Unit D was deposited. This unit is found in all the investigated cores from Lake Vetlavatn, and based on the AMS radiocarbon dates the 'Jondal Event 2' lasted for about 100 years. The TP-ELA lowering adjusted for land uplift is estimated to be 220 m during the 'Jondal Event 2'.

#### **Glacier magnitude and TP-ELA c. 10,450-10,000 cal. yr BP**

After the 'Jondal Event 2', the glacier retreated behind the local upstream watershed, and Unit C with high LOI values and no glacier meltwater input was deposited in lake Vetlavatn and found in all cores. After an early maximum, the minerogenic input of coarse silt suggested to indicate paraglacial reworking shows a gradual decrease to a minimum during this time span. The TP-ELA lowering adjusted

for land uplift is estimated to be 150 m during the time span 10,450-10,000 cal. yr BP.

#### **Glacier magnitude and TP-ELA c. 10,000-9900 cal. yr BP (the 'Erdalen Event 1')**

Represented by Unit B characterized by low LOI values and a distinct increase in coarse silt in all three cores, the glacier again advanced across the local upstream watershed for a short period, and glacier meltwater was rerouted towards lake Vetlavatn. Based on the AMS radiocarbon dates, the episode is suggested to represent the 'Erdalen Event 1' at Jostedalbreen (Dahl *et al.*, 1997).

The TP-ELA lowering adjusted for land uplift is estimated to be 210 m during the 'Erdalen Event 1' at northern Folgefonna.

#### **Glacier magnitude and TP-ELA c. 9900-9000 cal. yr BP**

Unit A consists of homogenous gyttja with high LOI values in the upper part of all investigated cores, including the 'Little Ice Age' period in cores III and IV. This was also confirmed by a HTH-core covering the upper 50 cm of the sediments. Based on the marginal moraines, however, the 'Little Ice Age' advance was the largest glacier event at northern Folgefonna after the 'Erdalen Event 1'. All marginal moraines formed by northern Folgefonna during the time span from c. 9900-9000 cal. yr BP were therefore erased by the 'Little Ice Age' glacier(s). Based on lake Vassdalsvatn, however, northern Folgefonna existed during this time span and did not melt completely away before about 7200 cal. yr BP (see Bakke *et al.* in prep. for further discussion).

The TP-ELA lowering adjusted for land uplift is estimated to be 140 m during the time span from 9900-9000 cal. yr BP at northern Folgefonna.

### **Palaeoclimatic reconstruction**

A close exponential relationship between mean ablation-season temperature  $t$  ( $^{\circ}\text{C}$ ) (1 May–30 September) and mean solid winter precipitation  $A$  (m water equivalent) (1 October–30 April) at the ELA of Norwegian glaciers in maritime to continental climatic regimes has been demonstrated (Sissons, 1979; Ballantyne, 1990; Dahl *et al.*, 1997, and references therein):

$$A = 0.915 e^{0.339t} \quad (r^2 = 0.989, P < 0.0001) \quad (\text{Eq.1})$$

The 'Liestøl equation' implies that if either the winter precipitation or the ablation-season temperature at the ELA is known, the other factor can be calculated. It also implies that if the former ELA is known, it is possible to quantify how the winter precipitation has fluctuated if an independent proxy for mean ablation-season temperature is used in the calculation (Dahl and Nesje, 1996a; Dahl *et al.*, 1997).

#### **Mean Lateglacial and early-Holocene ablation-season temperatures**

Suggested to be representative for northern Folgefonna, reconstructed summer temperatures (July) from Kråkenes at Vågsøy, western Norway (Fig. 1) (Birks and Ammann, 2000), have been used in the calculations. As a cirque glacier existed in the catchment at Kråkenes during the Younger Dryas (Larsen *et al.*, 1984), glacier meltwater had an increasing chilling impact on lake temperature with increasing air temperature during summer time. As a consequence, water-living organisms (chironomids, cladoceras etc.) can not be used for reliable temperature reconstructions during this time span.

Hence, for the Lateglacial (Younger Dryas) the summer-temperature reconstruction based on terrestrial plant macrofossils (Birks and Ammann, 2000) has been used, whereas chironomids have been used for the early-Holocene part (c. 11,500-10,000 cal. yr BP) (Brooks and Birks, 2000). As chironomids are not analysed from 10,000-9,000 cal. yr BP, terrestrial plant macrofossils are used during this time span (Birks and Ammann, 2000). The compiled summer-temperature (July) reconstruction (Fig. 12) is suggested to be representative for the ablation-season temperature at northern Folgefonna during the Lateglacial and early-Holocene (Fig. 12). A suggested standard error for both proxies corresponding to  $\pm 1^{\circ}\text{C}$  has been included in the figure. As only negligible adjustments relative to sea level have occurred at Kråkenes since the Younger Dryas, the reconstructed temperature curve is not adjusted for land uplift (Svendsen and Mangerud, 1987).

#### **Mean Lateglacial and early-Holocene winter precipitation**

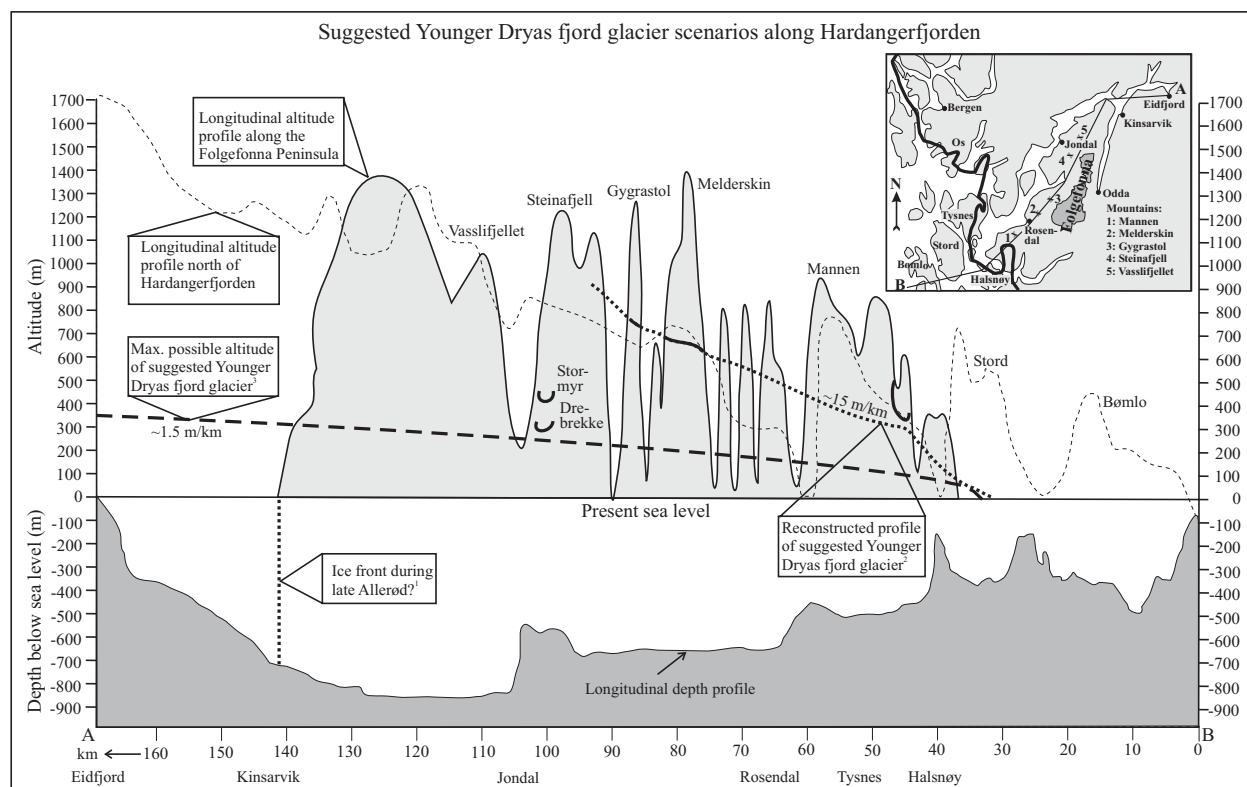
By combining the reconstructed TP-ELA curve adjusted for land uplift with the compiled summer-temperature curve from Kråkenes in Eq. 1, variations in mean solid winter precipitation have been quantified according to Dahl and Nesje (1996), and shown in Figure 13. The values are given as absolute variations relative to mean modern values and as variations in percent (mean 1961-1990 = 100%) (Klimaavdelingen, 1993b). Implications of the suggested standard error for the used summer temperature proxies of  $\pm 1^{\circ}\text{C}$  have been taken into account and included in Figure 13.

### **Discussion**

#### **Early deglaciation and age of low-altitude cirque glaciers**

The sedimentological investigations in lake Vetlavatn close to northern Folgefonna show a consistent AMS radiocarbon dated lithostratigraphy not covered by





**Figure 14** Suggested Younger Dryas fjord glacier scenarios along the Hardangerfjord modified from Follestad (1972). The upper dotted line shows the reconstructed Younger Dryas fjord glacier after Follestad (op. cit.), while the lower punctuated line marks the highest possible ice sheet/fjord glacier based on the Younger Dryas cirque glacier at Drebrekke in Jondal.

glaciers since c. 12,500 cal. yr BP (Fig. 7, Table 1). Bondevik and Mangerud (2002) dated the Younger Dryas glacier maximum to 11,700–11,600 cal. yr BP at Os just north of outer Hardangerfjorden (Fig. 1). If representative, this indicates that little glacier ice was produced at northern Folgefonna during the Younger Dryas. Combined with reconstructed sea-level fluctuations in inner parts of Hardangerfjorden showing a final deglaciation in late Allerød (Fig. 10) (Helle *et al.*, 1997), Jondal may have been deglaciated prior to the Younger Dryas. As Jondal is situated well inside the suggested margin of the Younger Dryas Scandinavian ice sheet at Halsnøy in outer parts of Hardangerfjorden (Fig. 1) (e.g. Follestad, 1972; Høltedahl, 1975; Mangerud, 2000), the area is important for the ice-sheet configuration in western Norway.

At Drebrekke in Jondal (Fig. 1), two sets of marginal moraines and indications of a third formed by a low-altitude cirque glacier are mapped down to 290 m above the present sea level (Figs. 1, 5 & 6). The farm at the marginal moraines at Drebrekke has a history back to the 15th century (Kolltveit, 1953), and together with lichen observations and Schmidt-hammer rebound values, this suggests that no glacier formed at Drebrekke during the 'Little Ice Age' (Figs. 4a & 4b). At Kråkenes on Vågsøy, western Norway, Larsen *et al.* (1984) mapped a similar setting of two marginal moraines and indications of a third formed by a Younger Dryas cirque glacier close to sea level.

The vertical difference between the modern TPW-ELA of ~1465 m at northern Folgefonna and the

TPW-ELA of the lowest active local cirque glacier is about 425 m. If both ELAs are adjusted for a glacio-isostatic land uplift of 100 m (Fig. 1), the difference between the Younger Dryas TPW-ELA of 1120 m at northern Folgefonna and the reconstructed TPW-ELA of 305 m asl. at Drebrekke, is 815 m.

The compiled Lateglacial and early-Holocene summer-temperature record from Kråkenes (Fig. 12) (Birks and Ammann, 2000; Brooks and Birks, 2000) indicates that no period after the Younger Dryas-Holocene transition had mean ablation-season temperatures colder than 9–10°C at the reconstructed TPW-ELA at Drebrekke. This is more than 4°C warmer than at any glacier ELA in southern Norway at present, and the estimated increases in regional winter precipitation as snow based on variations in the TPW-ELA of northern Folgefonna was not enough to compensate for the high summer temperatures during the early-Holocene (Fig. 13).

Derived from the above discussion, the low-altitude cirque glaciers at Drebrekke and Stormyr in Jondal are suggested to have formed during the Younger Dryas. The regional implications for the reconstructed Scandinavian ice sheet in Hardanger as summarized by Mangerud (2000) are shown in Figure 14. The reconstructed northern Folgefonna and the cirque glacier at Drebrekke during the Younger Dryas demonstrate that Hardangerfjorden most likely was deglaciated prior to the Younger Dryas, and that the production of glacier ice on surrounding mountains/plateaux was too restricted to fill up the wide and deep fjord during this event.

This strongly supports the reconstructed sea-level fluctuations in inner parts of Hardangerfjorden indicating a final deglaciation in late Allerød (Helle *et al.*, 1997). The suggested margin of the Younger Dryas Scandinavian ice sheet at Halsnøy in outer Hardangerfjorden (e.g. Follestad, 1972; Holtedahl, 1975; Mangerud, 2000) is poorly dated, and it may be older and/or have a local origin south of the main fjord. It also implies that the glacier advance dated by Bondevik and Mangerud (2002) to late Younger Dryas at Os just north of outer Hardangerfjorden most likely had a local origin in the Gullfjellet area on the Bergen Peninsula. As a consequence, the extent of the Younger Dryas glacier(s) in the Hardanger region was much less than previously suggested (e.g. Mangerud, 2000), and a major revision with implications for the Younger Dryas Scandinavian ice sheet in western Norway is therefore required.

#### Climate-induced early-Holocene glacier advances

After the Younger Dryas, the rapid retreat of the fjord glaciers has been associated with calving (e.g. Holtedahl, 1975; Andersen, 1980; Andersen *et al.*, 1995). Hence, and commonly referred to as Preboreal stages (e.g. Nesje and Dahl, 1993), fjord glaciers are suggested to have either readvanced or at least halted their general retreat as a response to steep and dynamically unstable glacier profiles. When the glacier became grounded and more dynamically stable on rock thresholds where the fjords become shallower, and/or where the valleys/fjords are relatively narrow, the glaciers formed frontal deposits as a response to the steep profiles (Kjenstad and Sollid, 1982; Sollid and Reite, 1983; Anda, 1984; Rye *et al.*, 1987).

The first direct evidence of climate-induced glacier advances after the Younger Dryas have previously been pre- 'Little Ice Age' moraines from the Erdalen Event readvance(s) at Jostedalsbreen, Grovabreen, Hardangerjøkulen and western Jotunheimen (see Dahl *et al.*, 2002, and references therein). By using rerouting of glacier meltwater across a local watershed in front of Nigardsbreen, firm evidence for a two-phase Erdalen Event associated with two readvances was achieved at this glacier (Dahl *et al.*, op. cit.). The first Erdalen Event readvance took place between 10,100 and 10,050 cal. yr BP, while the second occurred close to 9700 cal. yr BP.

Based on rerouting of glacier meltwater across a local watershed in front of northern Folgefonna, three early-Holocene climate-induced glacier readvances are inferred from a consistent AMS radiocarbon dated lithostratigraphy from lake Vetlavatn (Fig. 7, Table 1). The first named 'Jondal Event 1' took place c. 11,150-11,050 cal. yr BP within the temperature drop during the PBO (Björck *et al.*, 1997). The second named 'Jondal Event 2' occurred c. 10,550-10,450 cal. yr BP, while the third is suggested to be an equivalent

to the 'Erdalen Event 1' as described by Dahl *et al.* (2002) and took place c. 10,000-9000 cal. yr BP. This implies that especially early 'Preboreal stages' to some extent may have been influenced by the climatic induced readvances at high-lying glaciers during Jondal Event 1 and 2.

## Comparison with ELA and winter precipitation records in western Norway

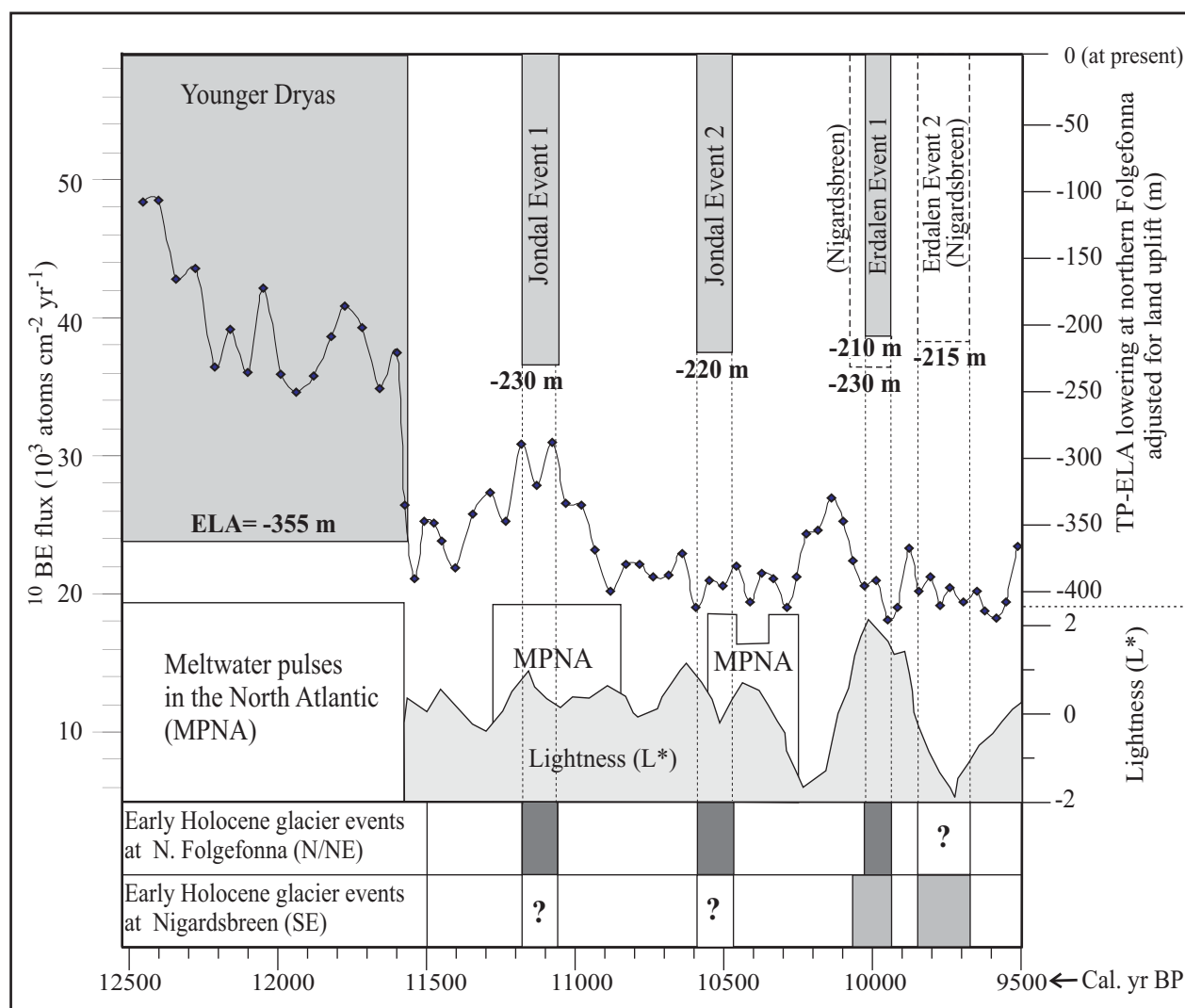
### The Younger Dryas

Only few records estimating the ELA lowering during the Younger Dryas are available from western Norway. Using modern glaciers further inland as a reference, Follestad (1972) estimated the ELA lowering of reconstructed glaciers of suggested Younger Dryas age to be 350 to 400 m at the south-western Folgefonna Peninsula. Based on results in the present paper, however, some of the reconstructed glaciers used by Follestad (1972) may be older.

Based on a reconstructed Younger Dryas cirque glacier at Kråkenes on Vågsøy in outer Nordfjord, Larsen *et al.* (1984) estimated the ELA lowering to about 700 m. As no glaciers exist in a similar setting in outer Nordfjord at present, an extrapolation of the modern ELA based on glaciers further inland (Liestøl, 1967) was used in the estimation. By suggesting the winter precipitation as snow to be similar to the present values, Larsen *et al.* (op. cit.) attributed the ELA lowering of 700 m to be the result of a drop in summer temperature of c. 4.2°C only. However, the present mean winter precipitation at Kråkenes is much less than some few tens of kilometres further inland, and makes the extrapolation of the modern ELA towards Kråkenes unlikely. Because of local topographic conditions, it is also difficult to compare an ELA lowering based on a reconstructed TPW-ELA of a cirque glacier at Kråkenes with the reconstructed TP-ELA lowering at northern Folgefonna.

By using the ELA of modern glaciers in inner Nordfjord as a reference, Fareth (1987) and Rye *et al.* (1987) estimated the ELA lowering during the Younger Dryas to be 400-500 m in this region. Based on the modern TP-ELA of a small plateau glacier at Storeloga close to Innvik, inner Nordfjord, Dahl and Nesje (1992) estimated the ELA lowering during the Younger Dryas to be about 500 m. All estimates on the Younger Dryas ELA lowering using modern glaciers in inner Nordfjord as a reference are in the same order as the estimate of 335 m from northern Folgefonna (Fig. 15).

Based on a similar approach as at northern Folgefonna, the estimated winter precipitation as snow was estimated to be less than 60 % compared to modern values in inner Nordfjord (Dahl and Nesje, 1992). This is wetter than the similar estimate of about 30 per cent during the suggested Younger Dryas at northern Folgefonna (Fig. 13), and may be explained



**Figure 15** Glacier events at northern Folgefonna and at Nigardsbreen (Dahl *et al.*, 2002) with corresponding ELA lowering estimates to the right, meltwater pulses in the North Atlantic (Clark *et al.*, 2001), lightness (Chapman and Shackleton, 2000) with scale to the right and atmospheric beryllium fluxes measured in the GRIP ice core (Björck *et al.*, 2001, and references therein) with scale to the left. The timing of the early-Holocene glacier events at northern Folgefonna and at Nigardsbreen are shown in the lower column.

by a shorter distance from Nordfjord to a suggested seasonally ice-free North Atlantic (e.g. Koc Karpuz and Jansen, 1992).

#### 'Jondal Event 1'/PBO

The glacier readvance during the 'Jondal Event 1'/PBO (Björck *et al.*, 1997) with a TP-ELA lowering of about 230 m is primarily suggested to be the result of somewhat lower summer temperatures for about 100 years (Figs. 12, 13 and 15) (Brooks and Birks, 2000). No other climate induced glacier advances/readvances has been demonstrated from western Norway during the PBO.

#### 'Jondal Event 2'

This climate-induced glacier readvance with a TP-ELA lowering of about 220 m is suggested to be the result of a marked increase in winter precipitation (Figs. 13 and 15). 'Jondal Event 2' may be contemporaneous with cirque glaciation recorded by Eikeland (Eikeland, 1991) at Sunnmøre, and it is suggested to be the first Holocene climate-induced glacier readvance/advance initiated by an increase in winter precipitation.

#### 'Erdalen Event 1'

The TP-ELA lowering of 210 m at northern Folgefonna during the suggested 'Erdalen Event 1' is primarily suggested to be the result of a marked increase in winter precipitation (Figs. 13 and 15).

This is in agreement with the conclusions by Dahl *et al.* (2002) from Nigardsbreen, a southeastern outlet glacier from Jostedalbreen. Compared with the estimates by Dahl and Nesje (1996), the calculated winter precipitation values during the 'Erdalen Event 1' may have been among the highest during the entire Holocene.

#### 'Erdalen Event 2'

No 'Erdalen Event 2' has been recorded at northern Folgefonna. At Nigardsbreen with a south-easterly aspect, however, this glacier (re)advance may primarily have been caused by a marked drop in summer temperature which reactivated already existing ice masses (Dahl *et al.*, 2002). Primarily because of dating uncertainties it is not obvious whereas other sites have recorded the first or the second Erdalen Event glacier readvance (e.g. Dahl

and Nesje, 1992; 1996a; Matthews *et al.*, 2000; Nesje *et al.*, 2001b). Why the 'Erdalen Event 2' is not recorded at northern Folgefonna is therefore poorly understood. However, aspect of the investigated outlet glacier(s) at northern Folgefonna and lack of a sensitive site may be of importance.

## Atmosphere-ocean interaction during the Lateglacial and early-Holocene

Suggested to be deglaciated prior to the Younger Dryas and sensitive to northern Folgefonna at present, Vetlavatn is the first Scandinavian record demonstrating climate-induced glacier fluctuations during the Lateglacial/early-Holocene transition. As the transition from glacial to interglacial conditions is not well understood, the atmosphere-ocean interaction leading to glacier advance and retreat in this region is of major scientific interest.

Among several hypotheses introduced to explain the apparent climate instability during the Lateglacial/early-Holocene transition, the most important are suggested to be:

- Large fresh-water outbursts into the North Atlantic may explain (some) abrupt climatic deteriorations during this time span (e.g. Broecker *et al.*, 1989; Broecker *et al.*, 1990; Clark *et al.*, 2001; Fisher *et al.*, 2002; Teller *et al.*, 2002).

- Variations in solar and/or geomagnetic forcing may explain the observed climate instability, and have been studied by use of  $^{10}\text{Be}$  and  $^{14}\text{C}$  records from the Greenland ice cores as a direct signal of changes in the production rates of the cosmogenic radionuclides (Björck *et al.*, 2001, and references therein).

- Fluctuations in the thermohaline circulation may strongly influence the climate in the North Atlantic region, and a lightness index based on deep-marine sediments has been used to study variations in the production of North Atlantic Deep Water (NADW) (Chapman and Shackleton, 2000).

In Figure 15, reconstructed records of fresh-water outbursts into the North Atlantic, solar and/or geomagnetic forcing and the production of NADW have been combined with Lateglacial and early-Holocene glacier fluctuations at northern Folgefonna.

Based on variations in  $^{10}\text{Be}$  from Greenland ice cores, the input of solar energy was low during the Younger Dryas (Björck *et al.*, 2001, and references therein), and there was a major meltwater pulse associated with this cold spell in the North Atlantic (e.g. Broecker *et al.*, 1989; Broecker *et al.*, 1990; Clark *et al.*, 2001; Fisher *et al.*, 2002; Teller *et al.*, 2002) (Fig. 15). Hence, the combined effect of this meltwater outburst and the low solar forcing apparently played a major role for a dry and cold climate in western Norway during the Younger Dryas.

Both the meltwater pulse and the solar forcing minimum ended at the termination of the Younger

Dryas. During 'Jondal Event 1'/PBO there was a new low in the input of solar energy, and another meltwater pulse occurred in the North Atlantic (Björck *et al.*, 1998; 2001, and references therein). Hence, the mechanisms behind the rather cold and dry 'Jondal Event 1' are suggested to be closely related to what happened during the Younger Dryas, but on a much smaller scale. Based on the lightness index (Chapman and Shackleton, 2000), a weak reduction in the thermohaline circulation in the North Atlantic took place just prior to the 'Jondal Event 1'/PBO (Fig. 15).

The initiation of the 'Jondal Event 2' took place in a period without meltwater pulses to the North Atlantic, and the glacier readvance terminated in the middle of one. Solar forcing based on  $^{10}\text{Be}$  was high during 'Jondal Event 2' (Björck *et al.*, 2001), and the thermohaline circulation was rather high just prior to this episode (Chapman and Shackleton, 2000). Hence, the high winter precipitation values estimated for the 'Jondal Event 2' may be the result of increased evaporation from warmer surface waters in the North Atlantic. The high winter precipitation values during this event (Fig. 13) may represent an early-Holocene occurrence of relatively mild, south westerly winds during winter time in western Norway. Based on modern instrumental records, these characteristics are associated with a positive North Atlantic Oscillation (NAO) mode in this region (e.g. Nesje *et al.*, 2000b, and references therein).

Predating the 'Jondal Event 2' and a North Atlantic meltwater pulse (Fig. 15), Björck *et al.* (2001) linked a cooling event at 10,300 cal. yr BP and one of the largest Holocene  $^{10}\text{Be}$  flux peaks. Contemporaneous, or somewhat prior to this  $^{10}\text{Be}$  flux peak, a marked reduction in the thermohaline circulation of the North Atlantic occurred (Chapman and Shackleton, 2000). No climate-induced glacier advance/readvance has so far been associated with this cooling event.

The 'Erdalen Event 1' is associated with a lack of meltwater pulses, strong solar forcing (Björck *et al.*, 2001) and a distinct increase in the thermohaline circulation of the North Atlantic (Chapman and Shackleton, 2000). Taking into account the high winter precipitation values to obtain the observed glacier readvance, the mechanisms behind 'Erdalen Event 1' may be analogous to 'Jondal Event 2'.

The 'Erdalen Event 2' took place in a period with no meltwater pulses, a rather high input of solar energy (Björck *et al.*, 2001) and a reduced thermohaline circulation in the North Atlantic (Chapman and Shackleton, 2000). No glacier readvance has been associated with 'Erdalen Event 2' at northern Folgefonna, but it has been recorded as distinct glacier readvance at Nigardsbreen, a south-easterly valley-outlet glacier to Jostedalbreen (Dahl *et al.*, 2002) (Fig. 15). The reason for this lack of evidence from northern Folgefonna may be a different aspect than Nigardsbreen.



## Conclusions

Based on the presented results and discussion the following conclusions and implications of local, regional and systematic importance are suggested:

1) The ice cap northern Folgefonna is suggested to have been isolated from the Scandinavian ice sheet during late Allerød and the entire Younger Dryas. This contradicts the previously presented model (e.g. Mangerud, 2000) where the Folgefonna Peninsula is suggested to be a main source of glacier ice to fill up Hardangerfjorden during this time span. The shift from a dry and cold Younger Dryas climate mode to the relatively warm and humid Holocene climate was rapid. As a consequence, northern Folgefonna is suggested to have shifted from a poly-thermal to a temperate temperature regime during this transition due to the sediments in lake Vetlavatn.

2) Based on the radiocarbon dated lithostratigraphy from lake Vetlavatn, lowering of the ELA, estimated Holocene winter precipitation values at northern Folgefonna and variations in Holocene summer temperatures, the low-altitude cirque glacier at Drebrekke between Hardangerfjorden and northern Folgefonna may have existed for the last time during the Younger Dryas. The existence of a cirque glacier in this topographical setting give a maximum altitude for the Scandinavian ice sheet during the suggested Younger Dryas which is well below the proposed model by Mangerud (2000) (Fig. 14).

3) The 'Jondal Event 1' is a climatic induced glacial readvance dated to 11,200 cal. yr BP, and with a TP-ELA lowering of c. 230 m. The event is suggested to be contemporaneous with the temperature drop during the Preboreal Oscillation (e.g. Björck *et al.*, 1997).

4) The climatically induced 'Jondal Event 2' occurred c. 10,550-10,350 cal. yr BP, and had a TP-ELA lowering of about 220 m. 'Jondal Event 2' is suggested to be the first Holocene glacier readvance/advance initiated by an increase in winter precipitation (Figs. 13 and 15).

5) An 'Erdalen Event 1' glacier readvance with a TP-ELA lowering of 210 m is suggested to have taken place at northern Folgefonna c. 10,000-9,850 cal. yr BP. The glacier readvance is primarily suggested to be the result of a marked increase in winter precipitation (Figs. 13 and 15), and the estimated values may have been among the highest during the entire Holocene.

6) The apparent lack of an 'Erdalen Event 2' at northern Folgefonna is poorly understood, but may be the result of a shift in the atmospheric circulation giving relatively more winter precipitation to glaciers with an aspect towards south-southeast.

7) Among several proposed hypotheses, the climate-induced early-Holocene glacier advances/readvances at northern Folgefonna are discussed in the context of large fresh-water outbursts to the

North Atlantic, variations in solar and/or geomagnetic forcing and by fluctuations in the thermohaline circulation of the North Atlantic (see discussion for references):

- 'Jondal Event 1' is closely related to a low the input of solar energy and to a meltwater pulse to the North Atlantic.

- 'Jondal Event 2' occurred in a period without meltwater pulses to the North Atlantic, solar input was high and the thermohaline circulation was strong. The high winter precipitation values estimated for this event may therefore be the result of increased evaporation from warmer surface waters in the North Atlantic. Hence, 'Jondal Event 2' may represent an early-Holocene occurrence of relatively warm, south-westerly winds during winter time in western Norway equivalent to a positive North Atlantic Oscillation (NAO) mode based on instrumental data in this region (e.g. Nesje *et al.*, 2000b, and references therein).

- The 'Erdalen Event 1' is, like the 'Jondal Event 2', associated with a lack of meltwater pulses, a strong solar forcing and a distinct increase in the thermohaline circulation. The winter precipitation values during this event may have been among the highest during the entire Holocene.

- The 'Erdalen Event 2' may be linked to a reduced thermohaline circulation. Whereas there is a coupling to a possible shift in the atmospheric circulation during this event is not known, but if such a shift took place this may explain the apparent lack of an 'Erdalen Event 2' at northern Folgefonna.

### Acknowledgement

This is a contribution from NORPEC, a strategic university program funded by the Norwegian Research Council (NFR). Specially thanks to Åsmund Bakke, Joachim Riis Simonsen, Øyvind Paasche and Jorun Seierstad for help during the field work and to Stig Monsen for help with the laboratory analysis.

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## Paper 3

Bakke, J., Lie, Ø., Nesje, A., Dahl, S. O. and Paasche, Ø. (submitted). A high-resolution Holocene glacier reconstruction based on physical sediment parameters from proglacial lakes at northern Folgefonna, western Norway. *The Holocene*





# A high-resolution Holocene glacier reconstruction based on physical sediment parameters from proglacial lakes at northern Folgefonna, western Norway

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Submitted to *The Holocene*

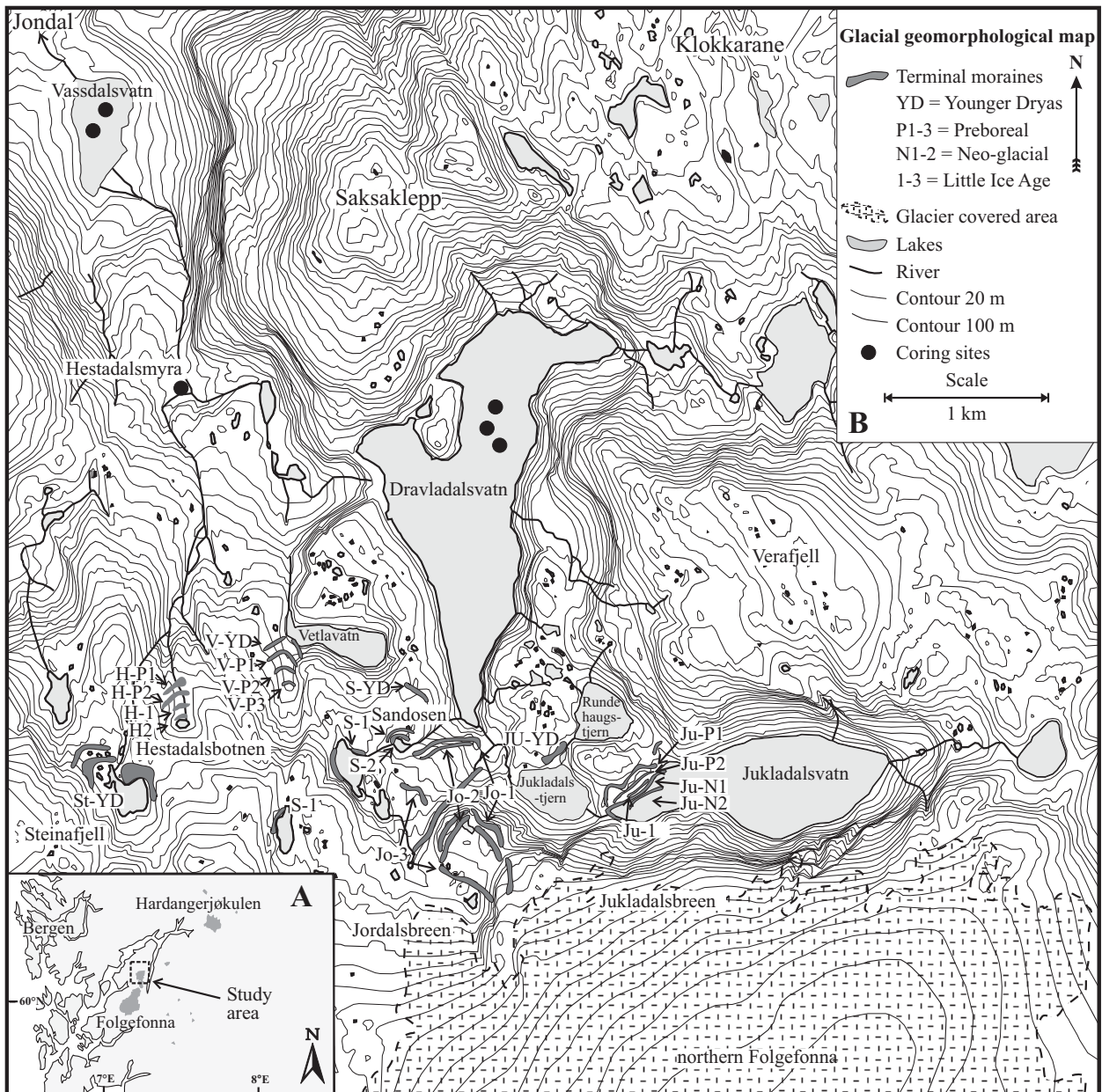
**Abstract:** The maritime plateau glacier northern Folgefonna in western Norway has a short (subdecadal) response time to climate shifts, and is therefore well suited for reconstructing high-resolution glacier fluctuations. The reconstruction is based upon physical sediment parameters in two proglacial lakes and a peat bog north of the ice cap. The record of glacier variations has been transferred into an equilibrium-line altitude (ELA) variation curve. Glaciers respond primarily to changes in summer temperature and winter precipitation. At present there is a high correlation between the North Atlantic Oscillation (NAO) index and measured (since the early 1960s) net mass-balance on maritime glaciers in western Norway ( $r = \sim 0.8$ ). Reconstructed glacier variations from the maritime western Norway can therefore be indicative of NAO-like weather modes during the Holocene. The early phase of mid-Holocene glacier growth was characterized by gradual expansion leading up to the first Subatlantic glacial event at 2300 cal. yr BP. The climate during the last 2200 years has been favourable for glacier growth at Folgefonna. High-amplitude shifts in ELA are interpreted as a result of unstable modes of the westerlies at the west coast of Norway, with significant changes in winter precipitation. Plotting the bulk density curve against the modelled glacier net mass-balance shows a remarkably similar pattern, where the maximum sediment yield was delayed by  $\sim 10$  years with respect to glacier net mass-balance. We here present a new method of reconstructing glacier variations in areas with low organic production. The approach is highly relevant in high-alpine and arctic regions where high-resolution reconstructions of former glacier variations are sparse.

**Key words:** Glacier fluctuations, North Atlantic Oscillation (NAO), bulk density, ELA reconstructions, Folgefonna, glacier mass-balance, Norway

## Introduction

Small plateau glaciers like Folgefonna in southern Norway (Figure 1) are ideal for studies of Holocene climate change because they respond rapidly to mass-balance perturbations (e.g., Dahl *et al.*, 2003), allowing the position of the equilibrium-line altitude (ELA) to reflect climate variability (e.g., Sutherland, 1984). Studies on modern Norwegian glaciers have shown that the sediment yield is positively correlated with glacier size (Roland and Haakensen, 1985). Measurements of the proportion of glacial material in proglacial lake sediments may provide continuous records of glacier fluctuations. The use of lake sediments in this context is widely used in Scandinavia (e.g., Karlén, 1976, 1981; Nesje *et al.*, 1991; 1995; 2000a; 2001; Matthews and Karlén, 1992; Dahl and Nesje, 1994; 1996; Snowball and Sandgren, 1996; Matthews *et al.*, 2000). Various approaches

related to proglacial sites uses a conceptual model of glacier-meltwater induced sedimentation in which the minerogenic (non-organic) component of the sediments is related to the presence of a glacier and its size in the catchment (e.g., Karlén, 1981; Leonard, 1985; Dahl and Nesje, 1994; Nesje *et al.*, 2000a; 2001; Dahl *et al.*, 2003). However, only a few studies have examined the physical properties of the sediments in detail and especially the minerogenic material produced by the glacier (Leonard, 1985; Souch, 1994; Rosqvist, 1995; Snowball and Sandgren, 1996; Matthews *et al.*, 2000; Nesje *et al.*, 2000a; 2001; Lie *et al.*, in press). The most common approach is to use the organic content [loss-on-ignition (LOI) and total organic carbon (TOC)] as an inverse indicator on inorganic deposition (op. cit.). In lakes with high minerogenic sedimentation and/or low organic production (<5%) this approach has its limitations since variations below this will not be detected. In



**Figure 1** (A) Map over southern Norway including glaciers (Folgefonna and Hardangerjøkulen) adapted from Østrem *et al.* (1988). (B) Detailed map over the moraine systems in the catchment north of the plateau glacier northern Folgefonna. Grey areas are terminal moraines mapped in the catchment. Black dots indicate coring sites and site for the sampling of the peat bog Hestadalsmyra. The lakes Dravladalsvatn and Vassdalsvatn, cored in this study, have inflow of meltwater from northern Folgefonna.

general, variations in physical sediment parameters in proglacial lakes are affected by sediment production-rates and discharge in the hydrological system, both factors mainly controlled by glacier size.

The climate at the west coast of Norway is influenced by advection of both warm water and air masses entering the NE Atlantic region, as well as position of the atmospheric polar front. The heat transport of the oceans gives the west coast of Norway large temperature anomalies (Broecker, 1991; Hopkins, 1991). Large temperature gradients generate cyclones crossing the North Atlantic region into and across Scandinavia. A close relationship between the winter weather and the North Atlantic Oscillation (NAO) index at the western part of Norway has been demonstrated (Hurrell, 1995; 2003;

Nesje *et al.*, 2000b). Through atmospheric general circulation models it is shown that the NAO probably is related to long-term trends in sea-surface temperatures (SST) (Feddersen, 2003; Hurrell *et al.*, 2003). It is assumed that higher winter precipitation in western Norway is related to stronger westerlies in the North Atlantic (and hence positive NAO weather modes). The mass-balance and hence size variations of maritime glaciers in western Norway may thus be indicative for long-term trends in the westerlies backwards in time. However, a high pressure field east of or, over Scandinavia gives a 'blocking' situation that forces the humid air masses either to the south or to the north of southwestern Norway (Shabbar *et al.*, 2001), hence complicating the interpretation of such atmospheric circulation



pattern.

Here we present a detailed, high-resolution reconstruction of the Holocene glacier variations of the maritime northern Folgefonna in western Norway. The main objectives in this paper are; (1) to reconstruct the Holocene glacial history of the plateau glacier Folgefonna at high temporal resolution, (2) to refine approaches for reconstruction of ELA variations using lake sediments with low organic/high minerogenic content, (3) to evaluate strengths and weaknesses of sediment parameters used to obtain high-resolution ELA reconstructions, (4) explain possible variability of atmospheric circulation modes during the Holocene, and (5) to compare the reconstructed glacial record from northern Folgefonna with measured net-mass-balance and modelled net mass-balance at Folgefonna.

## Study area

The ice cap northern Folgefonna covers an area of 23 km<sup>2</sup> and is the seventh largest glacier in Norway. It has a circular configuration and ranges from 1644 to 1200 m with a modern mean ELA of ~ 1465 m. Five major outlet glaciers flow from the ice cap; Jordalsbreen, Jukladalsbreen (Figure 1), Botnabreen, Dettébrea and Juklavassbreen. About 12 km<sup>2</sup> of the northern Folgefonna glacier drain northward and have a catchment that includes nine lakes.

The bedrock in the upper Jondal catchment consists mainly of acid meta-andesite, metadacite, quartzite, migmatite and migmatitic schist of Precambrian age (Sigmond, 1985; Askvik, 1995). The combination of acid rocks and the fact that the entire area lies above the tree line makes it a desolate landscape poor in both vegetation and superficial deposits. Except for some marginal moraines in front of the major outlet glaciers, there is only a sparse cover of colluvium and till in the area (Bakke, 1999). The absence of superficial deposits makes the area favourable for the use of lacustrine sediments from proglacial lakes, as the influence of paraglacial redeposition is minimized (Ballantyne and Benn, 1994; Ballantyne, 2002).

Based on a combination of two meteorological stations along Hardangerfjorden (Station no. 4949, Ullensvang Forsøksgård, 12 m a.s.l., 1962-1988; Station no. 5013, Omastrand, 1 m a.s.l., 1962-1990) (DNMI, 1993b), the present mean summer temperature ( $T_s$ ) from 1 May to 30 September is 12.7°C at sea level in Jondal. Using an environmental lapse rate of 0.6°C/100 m (e.g., Sutherland, 1984), this gives a mean  $T_s$  close to 4.0°C at the modern ELA (1465 m) of northern Folgefonna. At a local meteorological station about 11 km from the present glacier terminus of northern Folgefonna (Station no. 5696, Kvåle, 342 m a.s.l., 1961-1990) (DNMI, 1993a), the 1961-90 mean winter (1 October to 30 April)

precipitation ( $P_w$ ) was 1434 mm. Based on an empirical, exponential increase in winter precipitation with altitude of 8%/100 m in western Norway (Haakensen, 1989), the corresponding  $P_w$  at the ELA of northern Folgefonna is c. 3350 mm.

## Research approach and methods

The reconstruction of Holocene glacier fluctuations at northern Folgefonna is based on the following methods:

1. Detailed glacial-geomorphological mapping of the upper Jondal catchment, with special emphasis on former marginal moraines, glacier-meltwater channels and various ice-flow indicators (Bakke, 1999; paper 2).

2. Dating the 'Little Ice Age' glacial maximum and to sort out pre-LIA moraines. Lichenometry was used to determine relative age chronologies for marginal moraines (paper 2).

3. Study of lacustrine sediments in two proglacial lakes downstream from the glacier northern Folgefonna complemented by a peat bog stratigraphy.

4. The estimates of former glacier ELAs are based on observations of modern analogues and accumulated knowledge from previous studies (e.g., Andrews, 1975; Porter, 1975; Dahl *et al.*, 2003). Calculations at the glaciers are reconstructed using an accumulation area ratio (AAR) of 0.7 (e.g., Meierding, 1982; Dahl and Nesje, 1996). The moraine chronologies, and the reconstructed glaciers based on these, were used to calibrate the minerogenic signal in Dravladalsvatn. The calculation of the area distribution was carried out electronically using the vector-based GIS program MapInfo 6.0 on an N-50 map-datum.

## Fieldwork and laboratory analyses

A peat bog named Hestadalsmyra with bedrock thresholds in north and west, covers an area of 0.05 km<sup>2</sup> and is limited by the river from northern Folgefonna to the east and a small river from Hestadalsbotnen to the south (Figure 1). Whenever there is glacial activity in the cirque Hestadalsbotnen, the small river draining through the mire, transports and deposits glacially derived material into the site. The mire was cored by a 110 mm PVC tube that was hammered into the mire and then excavated. The tube was brought to the laboratory for radiocarbon dating and for magnetic susceptibility (MS) measurements.

Lake Dravladalsvatn (938 m) covers an area of 1.35 km<sup>2</sup> (Figure 2). The lake is situated in a glacially eroded bedrock basin with the longest axis (2.5 km) oriented north/south. This particular proglacial lake receives glacier-meltwater induced sediments whenever the glacier on the northern Folgefonna

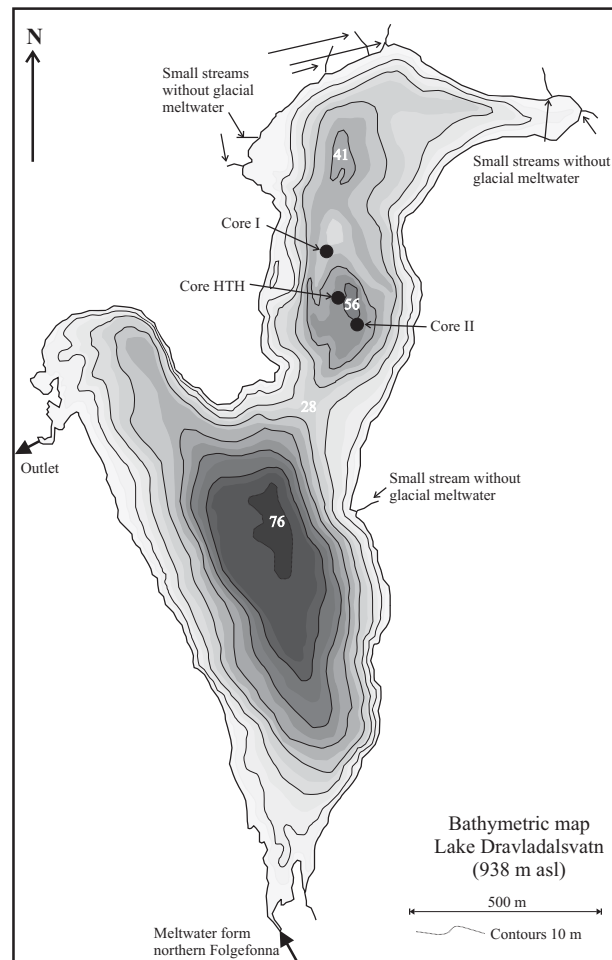
plateau is present. When there is no glacier present, gyttja dominates the sedimentation in the lake. The finest fractions of the meltwater-induced sediments are left in suspension in the water before settling. The distal, eastern basin, which was cored for this study, only receives the finest fractions of these sediments, as the basin is sheltered from the main river current. Lake Dravladalsvatn is the first basin to receive sediments from the glacier Jordalsbreen and the third to receive sediments from the glacier Jukladalsbreen. Since AD 1974 the lake level has been artificially raised due to production of hydro electric power.

Vassdalsvatn (490 m) covers an area of 0.17 km<sup>2</sup> (Figure 1) (bathymetry presented in paper 2), and is the seventh proglacial lake downstream from northern Folgefonna. The lake is located in a glacially eroded bedrock basin, and has at present input of glacier-meltwater induced sediments. Vassdalsvatn is suitable to record major flooding events in the catchment and to register whenever Folgefonna has been present.

Both proglacial lakes were cored using a modified piston corer taking up to 6 m long cores with diameter 110 mm (Nesje, 1992). In Dravladalsvatn an additional HTH gravity corer was used to retrieve the uppermost part of the lake sediments. The laboratory analyses for the cores from Dravladalsvatn included magnetic susceptibility (MS), weight loss-on-ignition (LOI) (Dean, 1974; Heiri *et al.*, 2001), dry bulk density (g/cm<sup>3</sup>) (DBD), water content, grain-size analysis using a Micromeretics Sedigraph 5100 (x-ray determination) (Sedigraph 5100, 1993). Grain-size statistics were performed by Gradistat 4.0 (Blott and Pye, 2001). The laboratory analyses for the two cores from Vassdalsvatn included LOI and MS.

'Sorting' and 'Mean' ( $\mu\text{m}$ ) have previously been used to distinguish between flood events and earthquake-triggered events (Arnaud *et al.*, 2002). The same approach has been adopted in this study to detect abrupt flooding events in Dravladalsvatn. Sorting (standard deviation of each sample) indicates the curve steepness of the cumulative grain-size distribution in a sample (higher values indicate poorer sorting), whereas 'mean' indicates the mean grain-size.

Twenty-five bulk samples and four macrofossil samples were AMS-dated from Dravladalsvatn and Vassdalsvatn. Terrestrial plant macrofossils for AMS radiocarbon dating were very sparse or absent in both lakes. The hard water effect/contamination is not regarded to cause any major problem since both lakes are located in acid Precambrian granite gneiss (Barnekow *et al.*, 1998; Lowe and Walker, 2000). The radiocarbon dates are shown in Table 1, and presented in the text as calibrated years before present, 1950 (cal. yr BP) according to INTCAL 98 (Stuvier *et al.*, 1998), if not otherwise stated.



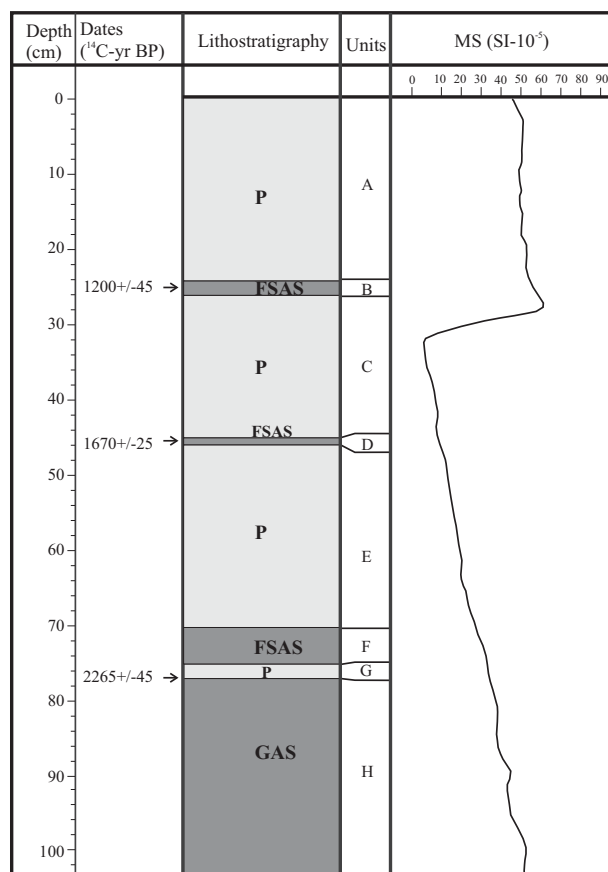
**Figure 2** Bathymetric map of Dravladalsvatn showing the location of the retrieved piston cores and the short gravity core (HTH). Note the bedrock threshold separating the two main basins with a 28 m deep sill. Inflow and outflow of glacial meltwater from northern Folgefonna are both in the western main part of the lake.

## Results

### Moraine chronology

Marginal moraines in front of the outlet glaciers from northern Folgefonna indicate up to eight successively smaller glacier advances/readvances (Figure 1). Beyond these moraines remnants of older glacier advances are present. The moraine chronology is not consistent around northern Folgefonna, which may be due to differences in aspect and slope at Jordalsbreen and Jukladalsbreen (paper 2).

The terminal moraines are marked with site name and numbers in Figure 1, and all moraines with numbers 1-3 were formed during the LIA (AD ~1750, ~1870, and ~1930 respectively). Calibrated against the 'Little Ice Age' moraines, Schmidt-hammer rebound values indicate that two marginal moraines may have formed c. 3000-1000 yrs BP (JU-N3 and JU-N2), whereas the rebound values for the remaining sets of marginal moraines (with lichen sizes over 200 mm) suggest a depositional age during the Lateglacial or early-Holocene (paper 2). Historical sources indicate river flooding events in Krossdalen



**Figure 3** Lithostratigraphy and MS in a core retrieved from the peat bog from Hestadalsmyra. Units B, D and F represent periods with glacial input to the peat bog (P = peat, FSAS = fine sand and silt, GAS = gravel and sand). The MS signal shows peculiar patterns, gradually decreasing up to unit B where it rises abruptly and remains high in the upper part.

(upper Jondal catchment) during the ‘Little Ice Age’ that caused damage to farmland (Kolltveit, 1953). Terminal moraines (Ju-P1, Ju-P2, Ju-N1, Ju-N2 and Ju-N3) lying north of Lake Jukladalsvatn demonstrate that Jukladalsbreen has crossed the valley several times during the Holocene. Glacial advances of Jukladalsbreen lead to the formation of a large lake to the east of the glacier that was catastrophically drained when the glacier retreated and/or the water pressure became higher than the ice pressure. During periods with extensive glaciers in Jukladalen, the drainage may have occurred randomly and to some extent also independent of climate, as ice thickness and water pressure controlled the water level.

In the ‘empty’ cirque Hestadalsbotnen there are four moraine ridges (Figure 1). P1 and P2 are assumed to be of Preboreal age and H-1 and H-2 were according to the lichen measurements deposited AD ~1870 and ~1750, respectively.

### Mire Hestadalsmyra

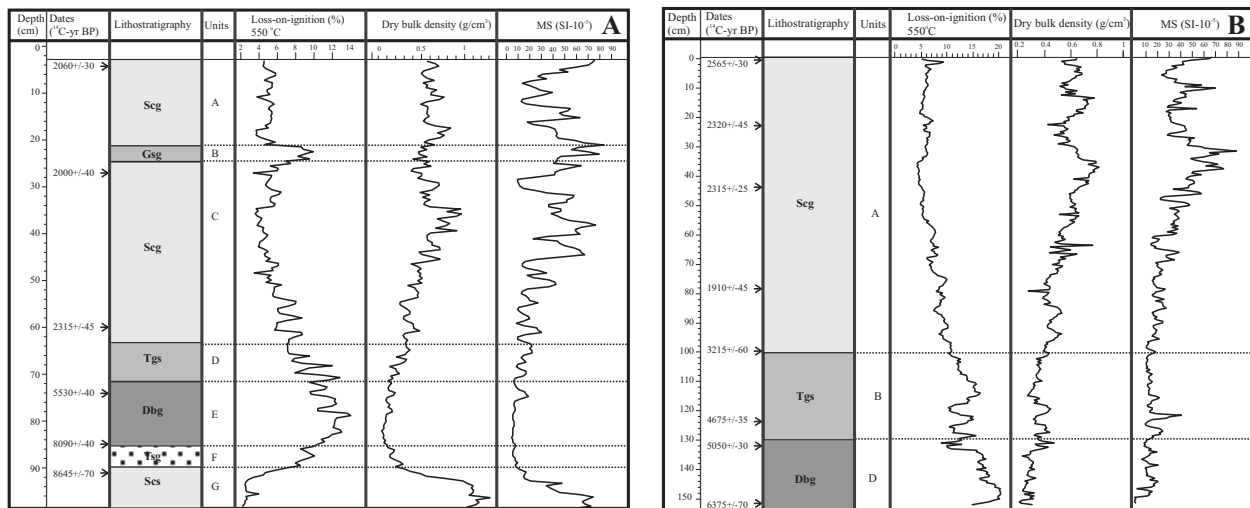
The lithostratigraphy in the peat bog has been divided into seven individual units (Figure 3). The lower unit (H) consists of gravel and sand with some plant macrofossils overlain by a short section of

humus (unit G). The lowermost layer of fine sand and silt (unit F) is 4 cm thick and dated at  $2265 \pm 45$   $^{14}\text{C}$  yr BP (T-3602) (for details regarding the radiocarbon dates, see Table 1). The upper boundary is gradual, whereas the lower boundary is sharp with a possible erosive contact to unit G. The next unit E, consists of homogeneous dark brown humus, whereas unit D is a 4 cm thick layer of fine sand and silt similar to unit F. A radiocarbon dating beneath the unit yielded an age of  $1670 \pm 25$   $^{14}\text{C}$  yr BP (T-13601). Unit C consists of humus similar to unit E, whereas unit B is a third 5 cm thick layer of fine sand and silt, radiocarbon dated in the upper part to  $1200 \pm 45$   $^{14}\text{C}$  yr BP (T-13600). The upper unit (unit A) consists of humus with grass on the top.

### Lake Dravladalsvatn

The interpreted lithologies from the individual cores are shown in Figure 4A-B. All three cores were taken in the deepest part of the inner basin of Dravladalsvatn (Figure 2). Core I was 97 cm long (Figure 4A) and the basal section consisted of a short sequence with suggested deglaciation sediments of grey silt and clay (unit G) below a gradually transition (unit F) into brown-dark gyttja (unit E). The transition between unit G and F is dated at  $8645 \pm 70$   $^{14}\text{C}$  yr BP (TUa-3629A). Above unit E, there was another transitional layer (unit D), going from dark brown gyttja to grey clay and silt. The basal part of unit E yielded an age of  $8090 \pm 40$   $^{14}\text{C}$  yr BP (Poz-3177) and of  $5530 \pm 40$   $^{14}\text{C}$  yr BP (Poz-3176) in the upper part. Unit C consisted of grey silt and clay with some lighter grey bands. A radiocarbon date in the lower part yielded an age of  $2315 \pm 45$   $^{14}\text{C}$  yr BP (TUa-3628A). The upper part of the unit was radiocarbon dated at  $2000 \pm 40$   $^{14}\text{C}$  yr BP (TUa-3627A). Unit B contained browner sediments dominated by silt and clay. The youngest unit A was similar to unit C, with a radiocarbon date at the top of  $2060 \pm 30$   $^{14}\text{C}$  yr BP (Poz-3175). The LOI pattern in the core showed higher values in the section dominated by gyttja with a decrease into unit C (Figure 4A). Through unit B the LOI values were higher than below, indicating higher organic content during deposition of this unit. DBD and MS are more-or-less in antiphase compared with the LOI values, with some higher variability in the MS, also showing, anomalous values throughout unit B.

Core II was 152 cm long (Figure 4B) and shows the same pattern as core I, except for the lowest part, which is missing in core II (units F and G in core I). The lower unit C consisted of dark brown gyttja with a radiocarbon date at the bottom yielding  $6375 \pm 70$   $^{14}\text{C}$  yr BP (TUa-3632A) and in the upper part  $5050 \pm 70$   $^{14}\text{C}$  yr BP (Poz-3256). Unit B is a transitional unit with a change from dark brown gyttja to grey silty gyttja. A radiocarbon date in the lower part yielded an age of  $4675 \pm 35$   $^{14}\text{C}$  yr BP (Poz-3179). Unit A consisted of grey clay and silt with an age in the lower part of  $3215 \pm 60$   $^{14}\text{C}$  yr BP (TUa-3631A). At 78 cm a



**Figure 4** (A) Lithostratigraphy, radiocarbon dates ( $^{14}\text{C}$ -yr BP), loss-on-ignition (% loss at 550 °C), DBD and MS measured at 0.5 cm interval in Dravladalsvatn core I (Scg = grey silt/clay, Dbg = dark brown gyttja, Tst = transition from silt towards brown gyttja, Scs = silty clay with stones, Tgs = transition from gyttja to silt/clay, Gsg = grey silty gyttja). (B) Lithostratigraphy, radiocarbon dates ( $^{14}\text{C}$ -yr BP), loss-on-ignition (% loss at 550 °C), DBD and MS measured at 0.5 cm interval for Dravladalsvatn core II (Scg = grey silt and clay, Tgs = transition from brown gyttja to silt/clay, Dgb = dark brown gyttja).

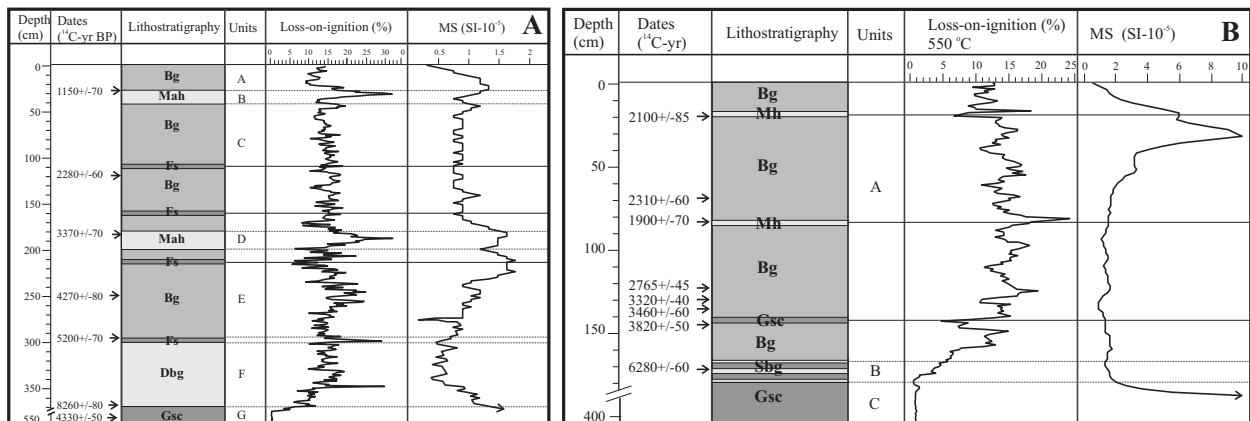
radiocarbon date yielded an age of 1910±45  $^{14}\text{C}$  yr BP (TUa-3630). In the upper part of unit A, three inverted radiocarbon dates were obtained, yielding 2315±25  $^{14}\text{C}$  yr BP (Poz-3798), 2320±45  $^{14}\text{C}$  yr BP (TUa-3640A) and 2565±30  $^{14}\text{C}$  yr BP (Poz-3178), respectively toward the top of the core.

A short 39 cm gravity core (HTH) from Lake Dravladalsvatn, consisted of a homogeneous section of silt and clay (Figure 7). The LOI was below 6% throughout the entire core. DBD and MS show high-frequency fluctuations. No radiocarbon dates have been obtained from this core.

### Lake Vassdalsvatn

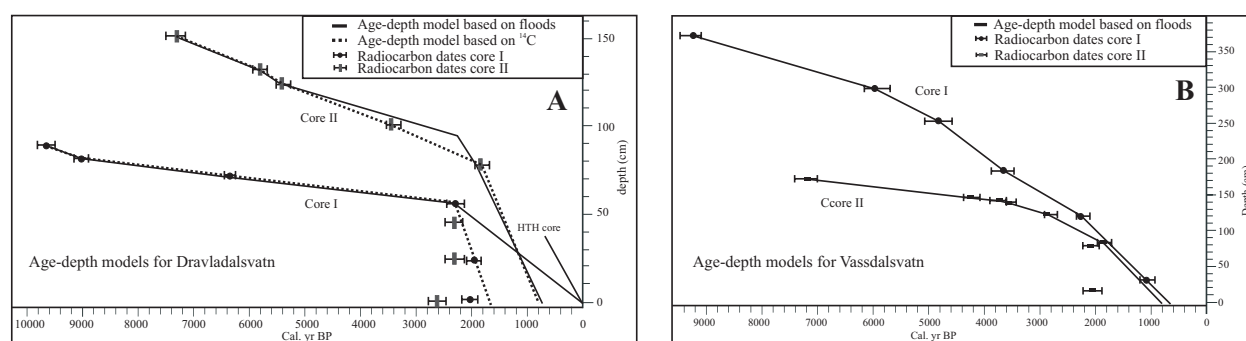
In Vassdalsvatn, the two cores showed remarkably different lithologies. Core I (550 cm) in Vassdalsvatn was retrieved in the central part, close to the main watercourse through the lake (Figures 1 and 5A). This core contained seven main units or lithological facies. Unit G consisted of a nearly 2-m long section of grey homogenous clay and silt, with a sharp transition to

the overlying unit. A radiocarbon date at the upper part of the unit yielded an age of 8260±80  $^{14}\text{C}$  yr BP (Beta-102936). Unit F consisted of dark brown gyttja that terminated in a layer of fine sand. The next (unit E) contained brown gyttja with a minor sand layer in the upper part. Two radiocarbon dates yielded ages of 5200±70  $^{14}\text{C}$  yr BP (Beta-102935) and 4270±80  $^{14}\text{C}$  yr BP (Beta-102933) in the lower and upper part of the unit, respectively. A layer of fine sand and macrofossils dominated unit D. A radiocarbon date of the layer yielded an age of 3370±70  $^{14}\text{C}$  yr BP (Beta-102932). The unit above (unit C) consisted of brown gyttja interlayered with two layers of fine sand. A radiocarbon date in the middle part of the unit yielded an age of 2280±60  $^{14}\text{C}$  yr BP (Beta-102931). A layer of fine sand and plant macrofossils (similar to unit D) dominated unit B. The transition to unit A, which contained brown gyttja, was radiocarbon dated at 1150±70  $^{14}\text{C}$  yr BP (Beta-102930). The LOI record shows an abrupt change from the lower minerogenic section (unit G) into unit F. The MS



**Figure 5** (A) Lithostratigraphy, radiocarbon dates ( $^{14}\text{C}$ -yr BP), loss-on-ignition (% loss at 550 °C) and MS measured at 2-cm intervals for Vassdalsvatn core I (Bg = brown gyttja, Mah = macrofossil and humus, Fs = fine sand and silt, Dbg = dark brown gyttja). (B) Lithostratigraphy, radiocarbon dates ( $^{14}\text{C}$ -yr BP), loss-on-ignition (% loss at 550 °C) and MS measured at 2-cm intervals for Vassdalsvatn core II (Bg = brown gyttja, Mh = macrofossil and humus, Gsc = grey clay and silt, Sbg = silty band with transition to gyttja).





**Figure 6** (A) Age-depth models for the three cores retrieved in Lake Dravladalsvatn. The solid line shows the linear interpolation model based on the radiocarbon dates in the each core. Dotted line shows the age-depth model correlated against the lithostratigraphy in Vassdalsvatn and the peat in Hestadalsmyra. Error bars shows 2 sigma. (B) Age-depth models fore the two cores retrieved in Lake Vassdalsvatn. The solid line shows the linear interpolation model based on the dates in the each core. Error bars shows 2 sigma.

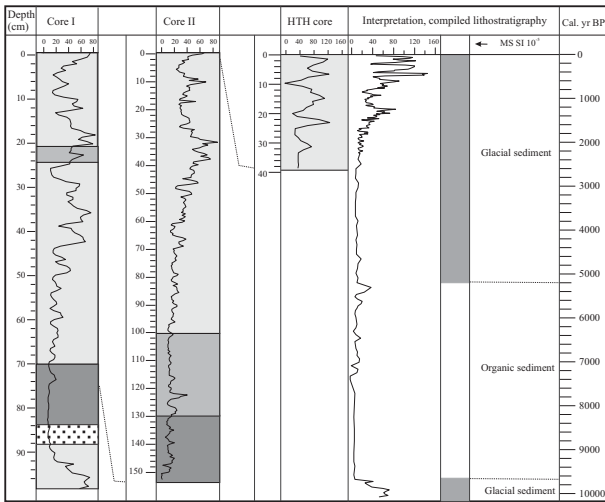
showed low values in unit F with a gradual rise in unit E and a marked peak in units D and B.

Core II (405 cm long) was retrieved in the eastern, more distal part of the lake (Figures 1 and 5B). This core was divided into three main units. The oldest (unit C) was 220 cm long and consisted of grey clay and silt. Unit B was a transitional unit from silty clay to gyttja that differed from the same transition in core I (unit G in core I). This unit contained three layers of grey silt and a radiocarbon dating in the middle of unit B yielded an age of  $6280 \pm 60$   $^{14}\text{C}$  yr BP (UtC-

6696). Unit A consisted of brown gyttja with two layers of macrofossils and fine sand. A layer of grey clay and silt was prominent in this bed and was radiocarbon dated at  $3820 \pm 50$   $^{14}\text{C}$  yr BP (UtC-6694) below and  $3460 \pm 60$   $^{14}\text{C}$  yr BP (UtC-6693) above. From this unit several radiocarbon dates have been obtained;  $3320 \pm 40$   $^{14}\text{C}$  yr BP (UtC-6692),  $2765 \pm 45$   $^{14}\text{C}$  yr BP (UtC-6691),  $1900 \pm 70$   $^{14}\text{C}$  yr BP (T-13608),  $2310 \pm 60$   $^{14}\text{C}$  yr BP (T-13788A), respectively (Figure 5B). The two layers with plant macrofossils and fine sand (unit D and B) were radiocarbon dated with

**Table 1** Radiocarbon dates obtained from the cores studied. When more than one calibrated intercept age is given, the median value is used.

Site	Lab nr.	Depth	Type of material	Radiocarbon age $\pm 1$ sigma	Intercept (cal. yr BP)	$\pm 1$ sigma (cal. yr BP)	$\pm 2$ sigma (cal. yr BP)
<b>Vassdalsvatn Core I</b>							
Vassdalsvatn I	Beta-102930	28-31	Gyttja	$1150 \pm 70$	1060	1170-970	1185-930
Vassdalsvatn I	Beta-102931	117-120	Gyttja	$2280 \pm 60$	2240	2350-2160	2360-2120
Vassdalsvatn I	Beta-102932	182-185	Gyttja	$3370 \pm 70$	3645	3690-3480	3830-3460
Vassdalsvatn I	Beta-102933	250-253	Gyttja	$4270 \pm 80$	4810	4965-4650	5045-4570
Vassdalsvatn I	Beta-102934	295-298	Gyttja	$5200 \pm 70$	5960	6170-5890	6170-5750
Vassdalsvatn I	Beta-102935	368-372	Gyttja	$8260 \pm 80$	9245	9415-9130	9430-9060
Vassdalsvatn I	Beta-102936	525-535	Gyttja	$4330 \pm 50$	4910	4965-4840	4990-4830
<b>Vassdalsvatn Core II</b>							
Vassdalsvatn II	T- 13607	19	Gyttja	$2100 \pm 85$	2050	2295-1970	2210-1890
Vassdalsvatn II	T-13608	83-84	Gyttja	$1900 \pm 70$	1840	1920-1735	1990-1690
Vassdalsvatn II	T- 13788A	77-79	Gyttja	$2310 \pm 60$	2080	2360-2160	2470-2150
Vassdalsvatn II	UtC-6691	123	Gyttja	$2765 \pm 45$	2860	2920-2785	2950-2775
Vassdalsvatn II	UtC-6692	138	Gyttja	$3319 \pm 40$	3550	3630-3475	3640-3465
Vassdalsvatn II	UtC-6693	142	Gyttja	$3460 \pm 60$	3720	3830-3640	3870-3570
Vassdalsvatn II	UtC-6694	147	Gyttja	$3820 \pm 50$	4250	4345-4100	4410-4090
Vassdalsvatn II	UtC-6695	171	Gyttja	$6280 \pm 60$	7205	7270-7030	7405-7005
<b>Hestadalsmyra</b>							
Hestadalsmyra	T-13600	26-28	Humus	$1200 \pm 45$	1120	1255-995	1280-965
Hestadalsmyra	T-13601	45,5-47,5	Humus	$1670 \pm 65$	1560	1690-1515	1715-1410
Hestadalsmyra	T-13602	75-77	Humus	$2265 \pm 60$	2250	2345-2160	2355-2145
<b>Dravladalsvatnet Core I</b>							
Dravladalsvatn I	Poz-3175	1	Macro fossil	$2060 \pm 30$	2030	2060-1990	2115-1950
Dravladalsvatn I	TUa-3627A	24	Gyttja	$2000 \pm 40$	1955	1990-1920	2045-1865
Dravladalsvatn I	TUa-3628A	57	Macro fossil	$2315 \pm 45$	2310	2355-2305	2465-2155
Dravladalsvatn I	Poz-3176	72	Macro fossil	$5530 \pm 40$	6340	6395-6290	6405-6280
Dravladalsvatn I	Poz-3177	82	Gyttja	$8090 \pm 40$	9055	9220-9000	9130-8980
Dravladalsvatn I	TUa-3629A	88	Gyttja	$8645 \pm 70$	9660	9690-9540	9800-9520
<b>Dravladalsvatnet Core II</b>							
Dravladalsvatn II	Poz-3178	1	Gyttja	$2565 \pm 30$	2620	2750-2550	2755-2490
Dravladalsvatn II	TUa-3640A	24	Gyttja	$2320 \pm 45$	2310	2360-2180	2465-2155
Dravladalsvatn II	Poz-3198	45	Gyttja	$2315 \pm 25$	2330	2350-2330	2355-2310
Dravladalsvatn II	TUa-3630	78	Macro fossil	$1910 \pm 45$	1840	1910-1745	1950-1725
Dravladalsvatn II	TUa-3631A	100	Gyttja	$3215 \pm 60$	3450	3475-3360	3575-3330
Dravladalsvatn II	Poz-3179	124	Gyttja	$4675 \pm 35$	5390	5465-5320	5470-5310
Dravladalsvatn II	Poz-3256	132	Gyttja	$5050 \pm 30$	5810	5890-5805	5900-5725
Dravladalsvatn II	TUa-3632A	151	Gyttja	$6375 \pm 70$	7305	7415-7250	7430-7180



**Figure 7** Compiled lithostratigraphy from Dravladalsvatn based on the three cores retrieved from the eastern distal basin. Correlation between the cores is done by using the age-depth models and the MS records.

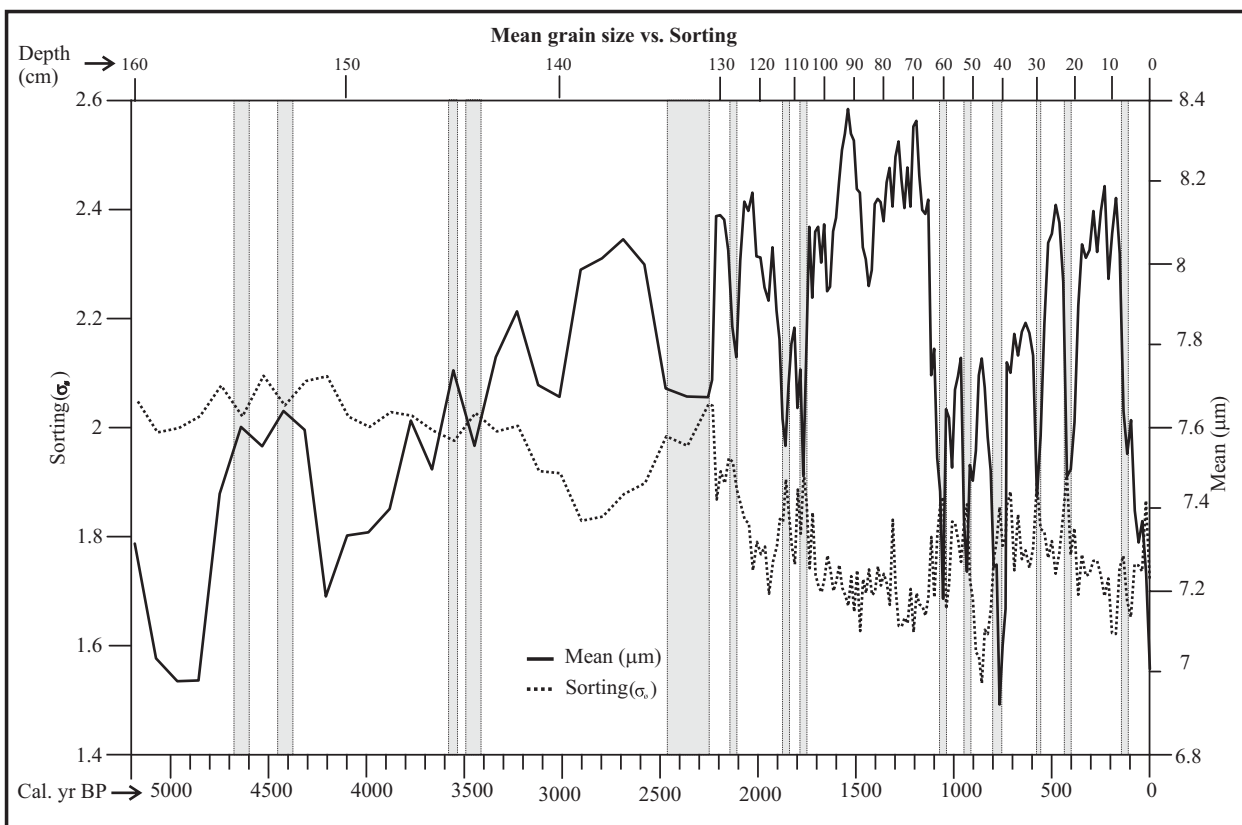
one sample below each unit. Unit D yielded an age of  $1900 \pm 70$   $^{14}\text{C}$  yr BP (T-13608), whereas unit B yielded an age of  $2100 \pm 85$   $^{14}\text{C}$  yr BP (T-13607).

### Age-depth models

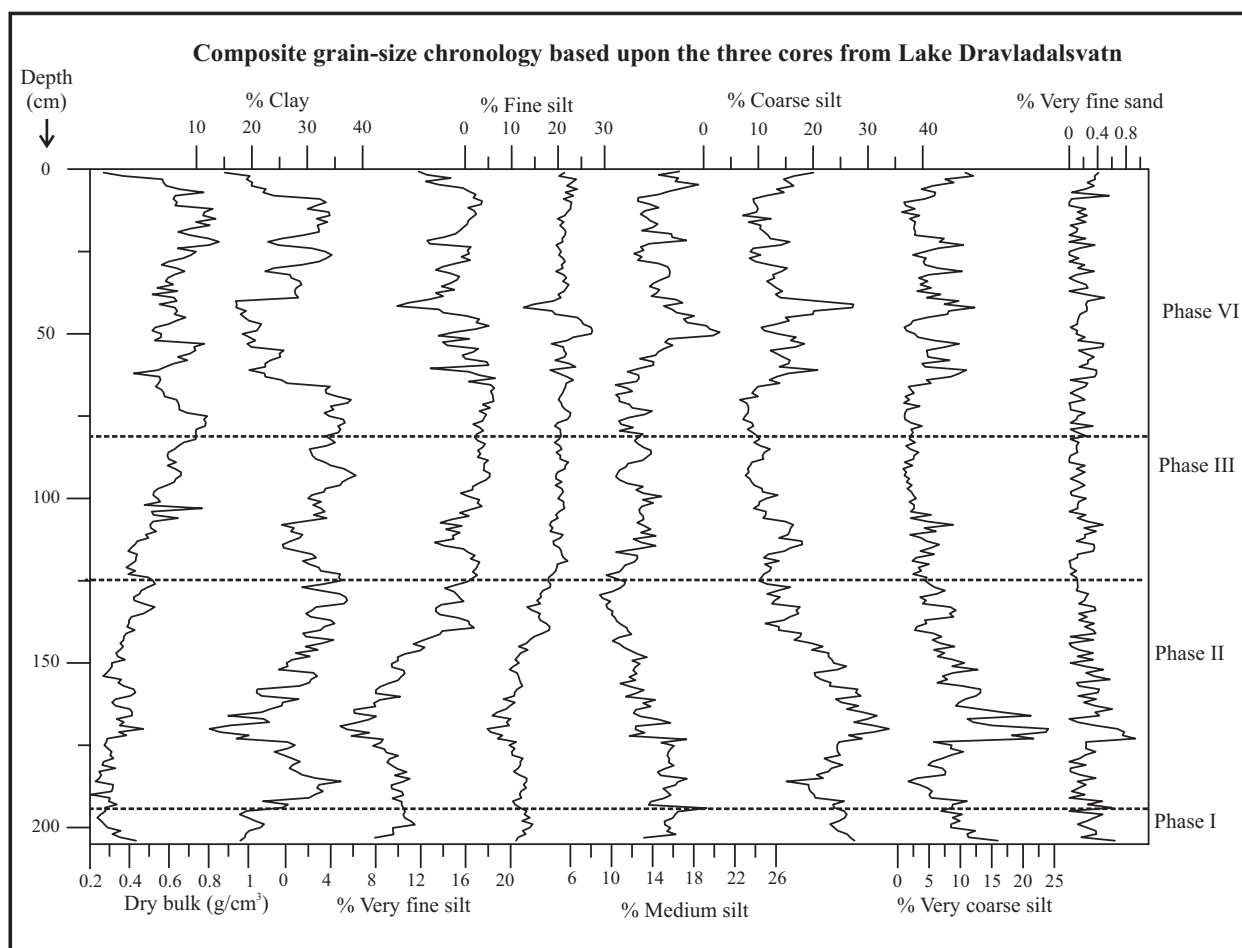
The age-depth models for Lake Dravladalsvatn and Lake Vassdalsvatn are shown in Figure 6A-B. Both models are constructed by linear interpolation between radiocarbon dates or between fixed points in the core lithologies. A major problem in the age-

depth modelling is that several of the radiocarbon dates during the last 2000 cal. yr BP are inverted (Figure 6). The inverted radiocarbon dates may be explained by resedimentation during floods or by increased input of meltwater from the glacier northern Folgefonna, because meltwater from glaciers can contain old carbon depending upon the age of the glacier ice that melts. AMS bulk dating can therefore be complicated during periods with large release of old water (melted glacier ice) as the amount of old carbon is increased. This is earlier regarded as a problem in deglaciation sediments with low organic content (Barnekow *et al.*, 1998; Lowe and Walker, 2000). However, this cannot be used as an explanation further downstream in Vassdalsvatn, where resedimentation is the most obvious reason for inverted radiocarbon dates.

By using flood events retrieved through analyses of sorting and mean as time markers, it was possible to correlate the lithologies between the sites. Periods with poorer sorting were interpreted as events with more input of sediments from suspension in Dravladalsvatn (Figure 8), interpreted as flooding events caused by glacier damming of the valley Jukladalen. Using this approach, four major flooding events were detected and correlated between Dravladalsvatn and Vassdalsvatn and used as time markers in the age-depth models. In Lake Vassdalsvatn, the flooding events were represented with layers of sand and in-wash of plant macrofossils. The sites belong to the same drainage system, and



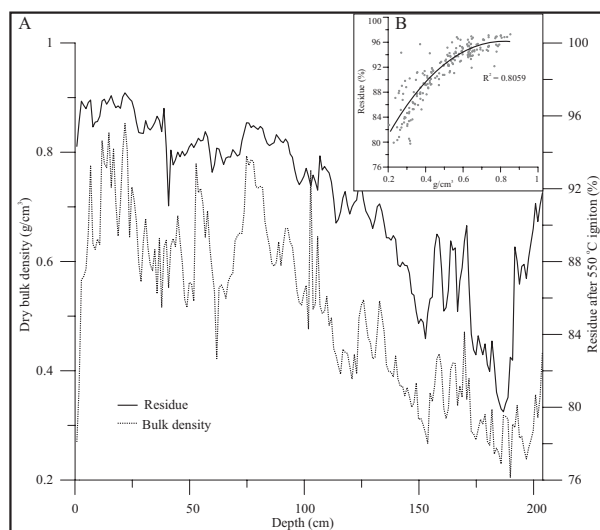
**Figure 8** Mean grain-size plotted against sorting (standard deviation in a sample). The upper axis shows depth (cm) in the compiled stratigraphy and the lower axis shows calendar year before present. As seen from the depth scale, there is a notable change in sedimentation rate around 130 cm. Higher 'sorting' values means poorer sorting of the sediments. Grey shaded areas show sorting anomalies.



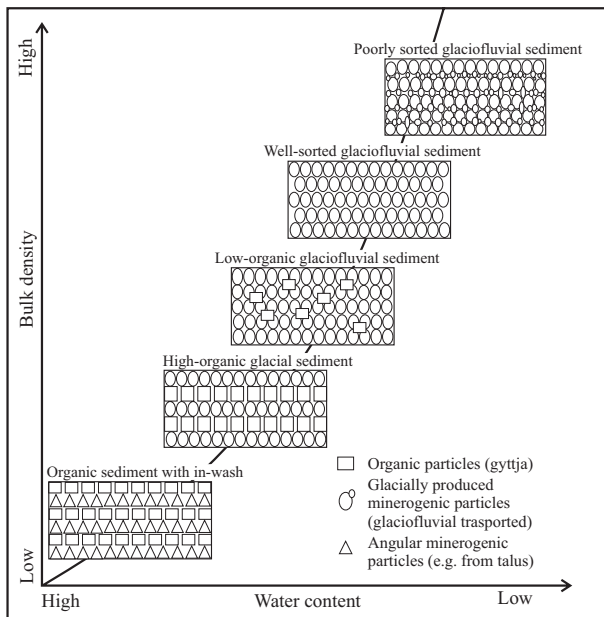
**Figure 9** Bulk density, clay, very fine silt, fine silt, medium silt, coarse silt, very coarse silt and very fine sand. The compiled lithostratigraphy is divided into four phases based on the presented parameters were they indicate notable changes in sedimentation environment.

these large floods should therefore be detectable in both lakes. In the peat bog Hestadalsmyra silt layers were interpreted to reflect floods or periods with glacial activity in the cirque Hestadalsbotnen. It is assumed that glacial activity in Hestadalsbotnen was mainly synchronous with periods of glacier growth at northern Folgefonna. The major flooding events occurred at ~3500 cal. yr BP (age from core I in Vassdalsvatn and core II in Dravladalsvatn), ~2250 cal. yr BP (age from core II in Dravladalsvatn, core II from Vassdalsvatn and core II in Dravladalsvatn), 1750 cal. yr BP (ages from mire Hestadalsmyra and core II in Dravladalsvatn) and ~1050 cal. yr BP (ages from peat bog Hestadalsmyra and core I from Vassdalsvatn). Beside these major floods there were several minor events observed in the sorting – mean record from Dravladalsvatn (Figure 8).

The short gravity core (HTH core) in Dravladalsvatn is suggested to overlap core II by three cm based on the same sedimentation rate as the upper part of this core (~10yrs/cm<sup>-1</sup>) (Figure 7).



**Figure 10** (A) Residue after 550°C ignition in % giving the minerogenic proportion after the organic content is removed. This parameter has traditionally been used as an indicator for inorganic sedimentation. Dotted line shows DBD. (B) Regression between residue (%) and DBD, showing a close relationship ( $r^2=0.8$ ) between the two parameters.



**Figure 11** Schematic figure explaining the relationship between bulk density and water content related to type of sediment. Angular minerogenic particles give higher porosity than rounded glacial-meltwater derived minerogenic particles. Lowest bulk density values are obtained from sediments dominated by gyttja and angular minerogenic particles.

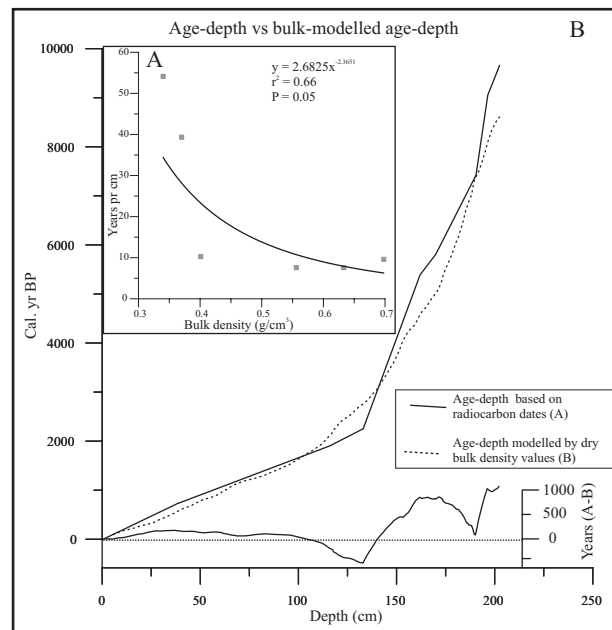
## Discussion

The composite stratigraphy from Dravladalsvatn is based on the age-depth model in Figure 7. Due to the large water depth (~70 m), it was difficult to retrieve the uppermost soft sediments. However, reliable correlations between the core lithologies were obtained by using MS and DBD records (Figure 7). Hence, by combining the three cores from Dravladalsvatn, a 202 cm long composite stratigraphy starting 9660 cal. yr BP (excluding the undated deglaciation section) was established.

The grain-size distribution in Dravladalsvatn is shown in Figure 9. Generally, there was a high input of coarse particles during time spans when the sediments were dominated by gyttja. The increase in very coarse silt is interpreted to be sediments from the surrounding catchment. When the LOI values decrease, a suggested glacial sediment consisting of clay and fine silt dominate the grain-size distribution. Negative correlations are evident between bulk density versus coarser fractions (very coarse silt and coarse silt) and positive correlations with the finer fractions. Based on these results the lithostratigraphy was divided into four phases (Figure 9).

MS show low values when gyttja dominates the sediments (phases III and II), whereas the MS values rise rapidly as the proportion of minerogenic sediments increased in phase IV. The high variability during phase I probably reflects varying influx of glacially derived sediments into the lake (Figure 7).

The glacial signal in Vassdalsvatn is suggested to be weaker because of longer transport length of the sediments compared to Dravladalsvatn. The lithostratigraphy in Vassdalsvatn was used to



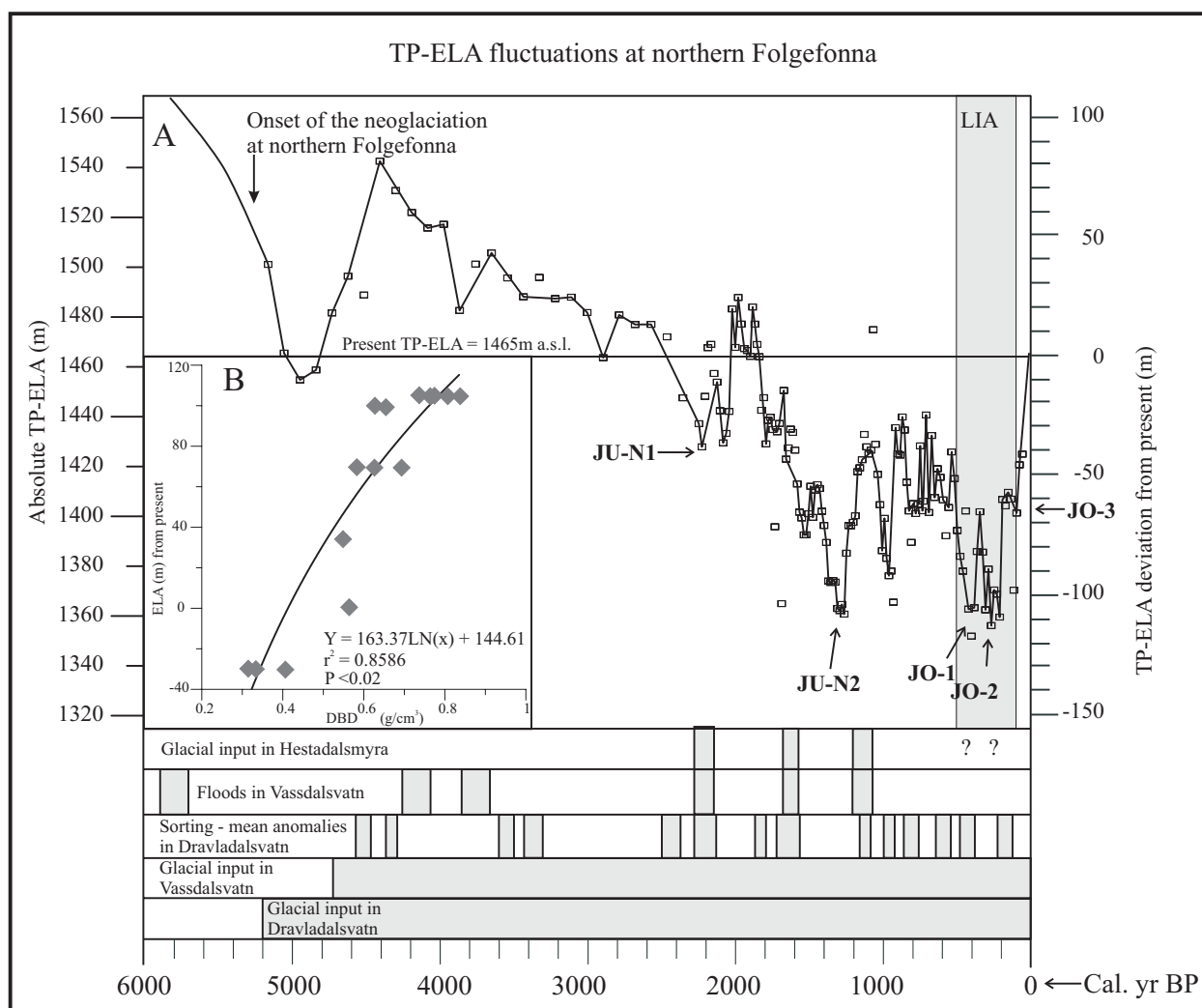
**Figure 12** (A) Regression between bulk density values and time resolution (years pr centimetre). The coefficient of determination (R-squared) shows high predictability due to sedimentation rate through the bulk values. (B) Modelled age-depth based on average bulk values between the radiocarbon dates in the compiled lithostratigraphy from Dravladalsvatn. The upper 100 cm is best reproduced in the model. The lower part of the figure shows the difference between the modelled age-depth and the age-depth model based on the radiocarbon dates.

complement Dravladalsvatn and for obtaining additional information regarding major flooding and glacier events in the catchment. The discrepancy in lithology between the two cores may be explained by the coring sites lying some distance apart.

## Loss-on-ignition, grain-size distribution and magnetic susceptibility

Loss-on-ignition is traditionally used as an inverse indicator for inorganic lake sedimentation. The approach is widely used to reconstruct glacier variations (e.g., Karlén, 1976; Karlén, 1981; Leonard, 1985; Nesje *et al.*, 1991; Dahl and Nesje, 1994; Rosqvist, 1995; Matthews *et al.*, 2000; Nesje *et al.*, 2000a; 2001). However, when the organic content is low (<5%), it is difficult to solve the amplitude of the glacial signal, as the signal-to-noise ratio becomes too low. This approach has therefore limitations in lakes in high alpine and arctic areas. In Dravladalsvatn, the LOI values were below 5% during periods when northern Folgefonna was at its present size (Figure 10). It is therefore difficult to obtain a continuous ELA reconstruction based on the LOI as an inverse indicator of the minerogenic sedimentation. Several physical sediment parameters describing the sediments produced by the glacier (e.g., bulk density and grain-size distribution) have therefore been taken into account. As seen in Figure 10, the overall patterns were reflected in both LOI and bulk density, but the bulk density record has a larger amplitude than the LOI record, in the minerogenic (low LOI) end of the spectrum, demonstrated by the exponential fit (Figure 10).





**Figure 13** (A) TP-ELA curve for Northern Folgefonna based on the bulk density record. The lower part of the figure is a summary of sources for validating the ELA record; the peat bog in Hestadalsmyra, floods in Vassdalsvatn, sorting anomalies in Lake Dravladalsvatn, glacial input in Vassdalsvatn and record of glacial input to Dravladalsvatn. (B) Regression between periods with known ELA (calibrated against the moraine chronology) and bulk density values. The regression model is used to transfer the bulk density record to a continuous TP-ELA curve.

### Bulk density as a proxy for glacier size

As the nature of glacial erosion is reflected by the supply of insoluble particles to a river system, analyses of physical properties of the glacial sediments may be a diagnostic parameter for variations in glacier size. Warm-based glaciers produce abundant clay-silt size fractions that are transported downstream to produce characteristic signatures in glacio-lacustrine sediments (e.g. Østrem, 1975). The use of grain-size variations have, however, not been widely used in this context. For calculating absolute amount of minerogenic sedimentation in a lake, it can be useful to flux correct the signal. A limitation for flux correction is commonly the number of radiocarbon dates. Another important factor concerning the grain-size distribution in proglacial lakes is that glaciers normally produce more than one dominating grain-size fraction. As seen from till studies, glaciers produce a composition of more-or-less all grain-sizes (e.g. Vorren, 1977). The glacial transport length and the size of the glacier do not seem to strongly

influence the grain-size distribution of glacial sediments (e.g. Jørgensen, 1977; Haldorsen, 1981, 1983). The grain-size variations in proglacial lakes are therefore mainly reflecting changes in fluvial and lacustrine systems. As the transport and sedimentation in fluvial systems are closely related to Hjulström's diagram (Sundborg, 1956), high-energy streams deposit less fine grained sediments, and *vice versa*. In 'open-ended' lakes, the finest grain-sizes will be transported further downstream because of stronger currents and slow settling. In a small, almost closed sediment basin, the grain-size distribution will consist of all grain-sizes suitable for suspension (1-63  $\mu\text{m}$ ), commonly giving more sediments per time than an 'open-ended' lake basin.

Bulk density acts as an additive parameter on the inorganic sedimentation. By definition bulk density expresses the ratio of the mass of dry solids to the bulk volume of a sediment (Blake and Hartge, 1986). Commonly, this parameter defines how granular, fibrous and powdery materials pack or consolidate under a variety of conditions and can be used to

calculate the porosity of the sediment. Changes in flux and packing (reflected in grain-size composition) are probably the most important parameter in a proglacial lake (Webb and Orr, 1997). Organic sediments should potentially be reflected by the lowest bulk values, whereas the highest values are expected in sediments consisting of fine-grained poorly sorted minerogenic sediments (Figure 11). Most of the source material in a proglacial lake has the same glacially derived origin and source is therefore neglected as a controlling factor for the bulk density values. Furthermore, changes in fluvial transport length can be neglected, as this have not changed during the Holocene. Water content is a parameter strongly linked to the bulk density parameter, as water fills the pores and expresses the porosity of the sediment (Menounos, 1997). In core II from lake Dravladalsvatn this relationship is very strong ( $r^2 = 0.97$ ), explaining most of the water content variance by variance in bulk values.

DBD values are assumed to be affected by changes in sedimentation rates. For testing of this relationship, average DBD values between each radiocarbon-dated level were plotted against the sedimentation rate. Based on this, an exponential relationship was established (Figure 12). The weakness of the model is lack of 'fixed' points in the age-depth model, and that only six intervals were used for establishing the regression. However, the coefficient of determination ( $r^2$ ) is 0.66, which indicates a significant relationship between sedimentation rate and DBD. The regression-model successfully characterises the overall changes in sedimentation rates. Especially phase IV (last 2000 yrs) is well reproduced, whereas phases II (9500-5200 cal. yr BP) and III (5200-2300 cal. yr BP) were less accurately reproduced. This may be explained by sedimentation dominated by gyttja during this time span (Figure 7).

By using the regression model above it was possible to flux correct the grain-size data for each sample. In Dravladalsvatn, very fine and fine silt showed considerable rise during the last 2000 cal. yr BP, and these two fractions are positively correlated to DBD.

## ELA variations at northern Folgefonna

A relationship between grain-size variations, DBD and glacier size based on the analysis has been established (Figure 13). The moraines JU-N1, JU-N2, JO-1, JO-2 and JO-3 have been used to calibrate the ELA curve by a correlation between ELA and DBD. Periods with sorting anomalies (due to flooding) have been removed from the ELA reconstruction (Figure 13). The sedimentation rate increased significantly during the last 2000 years and is suggested to be an independent verification for increased glaciers size during this time span (Figure

12).

The moraines in front of Jukladalsbreen have a somewhat different history than those in front of Jordalsbreen. This glacier is more sensitive to short-lived changes in precipitation and temperature because of aspect and surface geometry. Hence, the glacial events that caused the formation of moraines JU-N1 and JU-N2 are assumed to be successively smaller. This implies that these moraines were out of phase with the moraine chronology of Jordalsbreen. The DBD record is nevertheless regarded as representative for Jordalsbreen, as most of the sediments in suspension from Jukladalsbreen probably are deposited in Jukladalstjørn and Jukladalsvatn. As seen from Figure 13, the glacial advance JU-N1 gave a relatively small change in the total glacier-covered area compared with JU-N2. A possible explanation for this is that a cooling caused the event and that Jukladalsbreen advanced more than Jordalsbreen, as the accumulation area for this glacier is located at higher altitude.

The ELA reconstruction has been divided into six phases:

1. Between 9600 and 5200 cal. yr BP the ELA at northern Folgefonna was above the highest mountain (> 1550m) and there was no glacier present in the catchment.

2. ELA dropped around 5200 cal. yr BP and the Folgefonna glacier was reformed after the 'thermal optimum'.

3. From ~4600 to ~2300 cal. yr BP there was a gradual build-up of northern Folgefonna towards its present size.

4. Around 2200 cal. yr BP there was a double short-lived glacier advance followed by a rapid rise in ELA (~1500 m) around 2000 cal. yr BP.

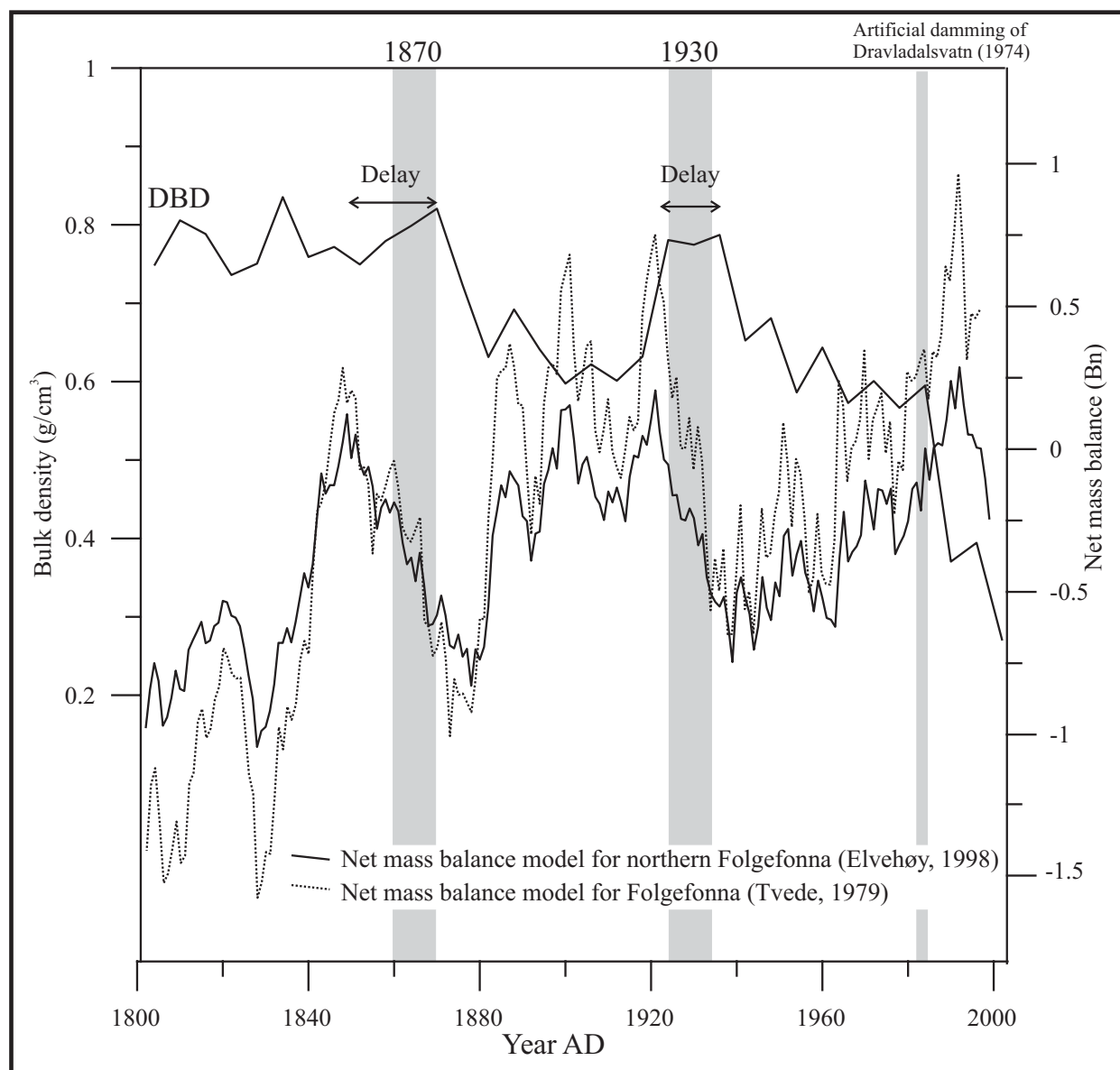
5. From 2000 until ~1400 cal. yr BP the ELA was lowered to 1360 m by a gradual build-up of the glacier size.

6. From 1200 cal. yr BP until present the ELA variations led to high-frequent changes in glacier size before a rather long period (from ~600 cal. yr BP until AD 1930) with large glaciers during the 'Little Ice Age'.

## Bulk density and net-balance modelling

DBD as a proxy for former glacier size is a new approach, and the validity is tested against net-mass-balance data for the last 200 years at Folgefonna (Figure 14). Tvede (1979) established some equations for modelling of the glacier mass-balance ( $B_w/B_s$  and  $B_n$ ) of Folgefonna based on temperature and precipitation from the Bergen meteorological station (st. no. 5054/56). The equation was later reformulated (Elvehøy, 1998) and established also for the northern part of Folgefonna:

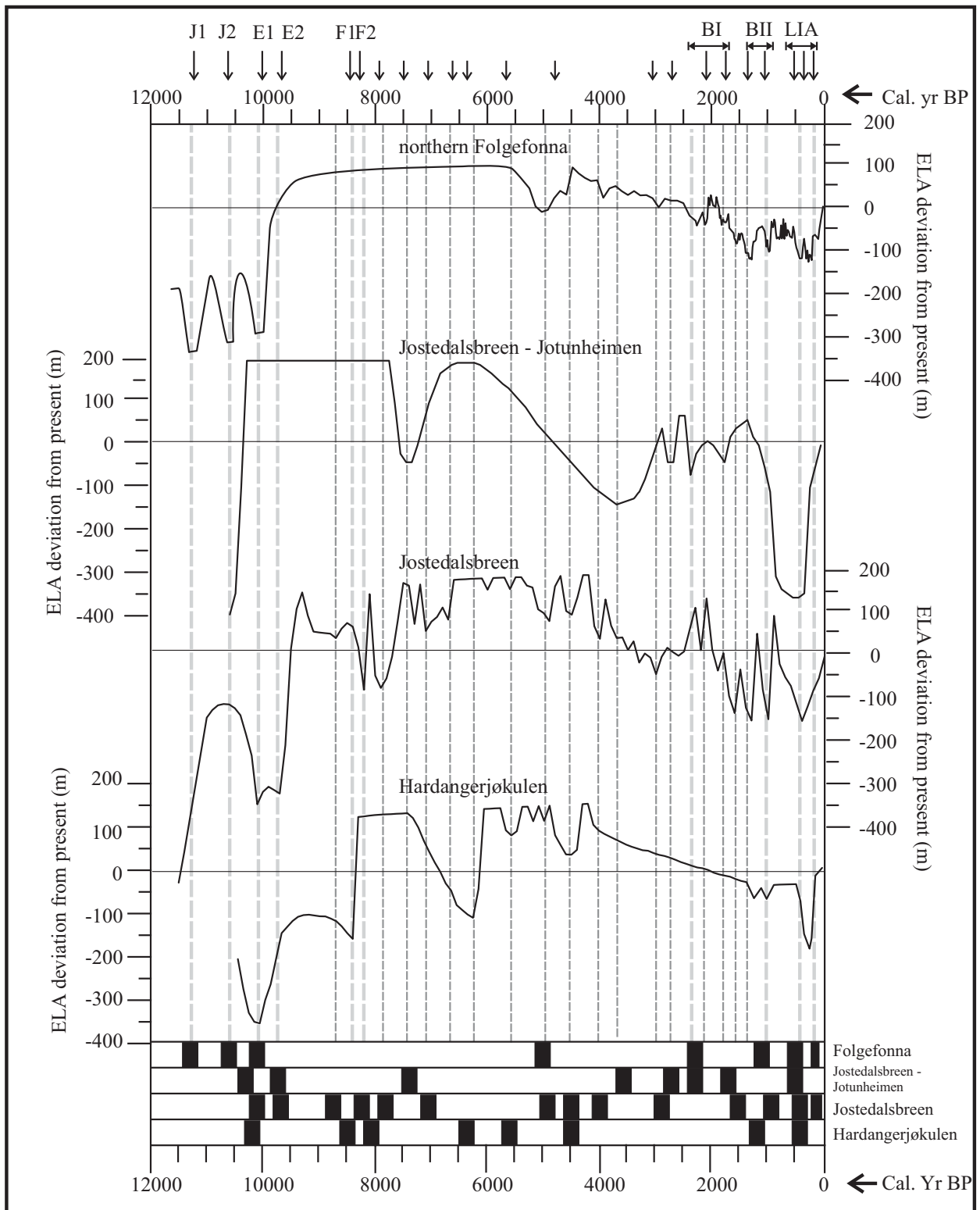
$$B_n = 444 + 2.16 * P - 54 * T_s,$$



**Figure 14** Dry bulk density (DBD) compared to two different glacier net mass-balance models for the Folgefonna and the Northern Folgefonna glaciers. The dotted line shows a model developed by Tvede (1979) for the southern part of the Folgefonna glacier, whereas the black line was produced for northern Folgefonna in this study. Both models are tuned against a glacier net mass-balance model and net balance measurements from AD 1963 to AD 1997 (Elvehøy, 1998) at Folgefonna. The reconstruction is based on temperature and precipitation records from Bergen and Ullensvang (Birkeland, 1951, DNMI, 1993a;b). Thick-grey shaded areas indicate periods with moraine formation.

where  $P$  is winter precipitation in Bergen (01.10-31.05) and  $T_3$  is average summer temperature (01.06-31.08). The equation gives high predictability compared to the net mass-balance from 1963-1997 ( $R^2 = 0.84$ ). In the reconstruction, temperature and precipitation records from Bergen back to AD 1841 (data from Meteorological Institute) were put into the equation, whereas a temperature record from Ullensvang was used from AD 1800 – 1840 (Birkeland, 1951). As there is a lack of precipitation records for this time span, a linear regression model between the January, February and March temperatures ( $r = 0.6$ ) to reconstruct the winter precipitation was used. Both models reproduce the AD 1870 (late LIA) glacier advance and the AD 1930

glacier advance, which correspond to moraines Jo-2 and Jo-3 at Jordalsbreen, respectively. The low  $B_n$  values from AD 1800 to AD 1840 are reflecting the retreat of the glacier after the AD 1750 glacier event, and may explain the high DBD values during the time span. Another interesting feature is that there exist lags in the bulk density record with  $\sim 10$  years from a change in net balance to increased bulk density values. This is suggested to reflect the response time of the glacier to mass-balance perturbations. Hence, the test is suggested to strengthen the link between the reconstructed ELA curve based on bulk density against modern instrumental and glacier mass-balance data.



**Figure 15** Compilation of some selected ELA reconstructions from southern Norway throughout the Holocene period (Paper 3 and 2, Matthews and Karlén, 1992, Nesje *et al.*, 2001, Dahl and Nesje, 1996). Upper panel with arrows shows glacial events (abrupt events of decadal to millennial duration) derived from the compiled glacial records. During the early-Holocene most of the events are named (J1 = Jondal Event 1, J2 = Jondal Event 2, E1 = Erdalen Event 1, E2 = Erdalen Event 2, F1 = Finse Event 1, F2 = Finse Event 2), whereas the mid-Holocene has several unnamed events. Late-Holocene events are named at Bøvertunsbreen (BI = Bøvertun Event 1 and BII = Bøvertun Event 2) (Matthews *et al.*, 2000).



## Implications of climatic importance from the northern Folgefonna record

The approach shown in this paper provides new methodologically tools for reconstruction of glacier variations from lake sediments with low organic content. The approach is appropriate in high alpine and arctic regions, where high-resolution reconstructions of former glacier variations from lake sediments are sparse.

In Figure 15, the reconstructed glacier variations at northern Folgefonna are compared with other studies from southern Norway. Based on the adopted ELA curve in this study, several implications follow from the temporal pattern of Holocene glacier variations at northern Folgefonna:

(1) In paper 2, the early-Holocene event-chronology is discussed on the basis of lake sediments from Vetlavatn, indicating three episodes of glacier advance subsequent to the Younger Dryas. The second Erdalen Event glacier readvance did not cross the threshold to Vetlavatn, and it is therefore not recognised as a glacial readvance at northern Folgefonna. However, the data from Dravladalsvatn indicate a rapid retreat of the glacier, as the initial gyttja-dominated sediments are dated to 9660 cal. yr BP. This is in accordance with the data from Matthews and Karlén (1992) in the Jostedalsbreen – Jotunheimen area. At Hardangerjøkulen and Jostedalsbreen the glaciers existed continuously until the end of the Finse Event (~8.2 ka cal. yr BP).

(2) The record from northern Folgefonna does not support glacier readvances during the double Finse Event, which were first recorded in a peat bog at Finse (Dahl and Nesje, 1994; 1996). The event has later been reproduced from other peat sections, proglacial and non-glacial lakes (Dahl and Nesje, 1996; Matthews *et al.*, 2000; Nesje *et al.*, 2000a; 2001; Nesje and Dahl, 2001). A possible explanation for this discrepancy may be the altitudinal range of the glacier. In Matthews and Karlén (1992), the importance of glacier altitude regarding temperature changes was examined, and they concluded that the highest-lying glaciers existed longer into the 'thermal optimum' than lower-lying glaciers. As the glaciation threshold due to the topography around northern Folgefonna is 1550 m a.s.l., the lowering of the ELA during the Finse Event may not have crossed the glaciation threshold or the altitude of instantaneous glacierization (AIG) (Lie *et al.*, 2003). During the maximum of the Finse Event at Hardangerjøkulen, the TP-ELA was 1580 m (Dahl and Nesje, 1996), which is above the highest part of the subglacial mountain plateau beneath northern Folgefonna. The plateau glacier Hardangerjøkulen lies 80 km to the northeast of northern Folgefonna, and the TP-ELA is therefore regarded as comparable.

(3) The first neoglaciation at northern Folgefonna started around 5200 cal. yr BP. This is different from the other reconstructions from southern Norway

where there were several shorter and longer glacial periods between 9660 and 5200 cal. yr BP. The reason for this is most likely the altitudinal range of northern Folgefonna. This is also indicated for the glacier Ålfotbreen that has a modern mean TP-ELA of ~1200 m. Here, the neoglaciation started ~2330±60 cal. yr BP by some smaller glacier events, before the glacier recovered and existed continuously from around 850 cal. yr BP (Nesje *et al.*, 1995).

(4) The first neoglaciation at northern Folgefonna has some notable consistent modes. The glacial advance from 5200 cal. yr BP until ~2300 cal. yr BP was a phase with gradually glacier growth. Such a gradual transition is known from other palaeoclimatic archives in the North Atlantic region, especially prominent in the reconstructed sea-surface temperatures (SST) at the Vøring Plateau (Calvo *et al.*, 2002). The SST record shows marked drops in temperature at 5400 and 2500 cal. yr BP, which correspond to marked changes in glacier size at northern Folgefonna. It is therefore assumed that the boundary conditions for glacier growth in southwestern Norway indicates a change in the atmospheric and oceanic conditions, rather than abrupt climatic changes as recorded during the early-Holocene (Dahl *et al.*, 2002; Paper 3). The high correlation against the SST reconstruction indicate that the ocean has a major forcing on the atmospheric circulation. Throughout the neoglaciation it is recognised a possible 1ky cyclicities in the ELA record at northern Folgefonna.

(5) The Holocene glacier record from northern Folgefonna indicates high-frequent changes in glacier size during the last 2300 years, with century to millennial-scale glacier expansions and some less extensive decadal glacier fluctuations. Of special interest are three relatively large glacial readvances dated at 2200, 1600 and 1050 cal. yr BP. Periods with glacier expansion is also recognised at Jostedalsbreen (Nesje *et al.*, 2001), Hardangerjøkulen (Dahl and Nesje, 1994) and at Bøvertunsbreen (Matthews *et al.*, 2000) in southern Norway during the same time span. From northern Norway, two late Holocene glacier readvances are recognised at Okstindan dated at 3000-2500 <sup>14</sup>C yr BP and 1250-1000 <sup>14</sup>C yr BP (Griffey and Worsley, 1978). At Okstindan, these glacial readvances were larger than during the "Little Ice Age". This is similar to the record of Jukladalsbreen at northern Folgefonna. Based on the wide geographical distribution of the late Holocene glacier advances, it is assumed that these high frequent climate shifts, leading to glacier expansion and decay, are representative for at least western Scandinavia. Folgefonna is a maritime glacier where ~80% of its modern  $B_n$  is forced by changes in  $B_w$ . Possible explanations for the changes around 2200 cal. yr BP may therefore be in the record of winter precipitation (providing NAO+ weather mode) that appears in a less stable mode than during the period from 5200 - 2200 cal. yr BP. Another explanation

maybe stronger effect of the Russian High, giving variable patterns of the westerlies and hence the precipitation pattern along the west coast of Norway (Hurrell, 1995; Shabbar *et al.*, 2001; Hurrell, 2003).

(6) The period termed 'the Medieval Warm Epoch' (MWE) is without any significant signature in the glacial record from northern Folgefonna. The MWE is referred to as the time interval between AD 800 and AD 1300 (Cronin *et al.*, 2003). It is apparent that the temperature record did not exceed the present temperature range (Crowley and Lowery, 2000; Bradley *et al.*, 2003). If a MWE temperature rise was followed by an increase in winter precipitation, the glacier at northern Folgefonna would have expanded. This is suggested from the ELA reconstruction in this study, as the time interval for the MWE includes both periods of glacier expansion and decay. A possible explanation for the glacier decay could be higher winter temperatures, which could give rain instead of snow at the glacier.

(7) The 'Little Ice Age' at northern Folgefonna had three periods of glacier growth peaking at AD ~1750, AD ~1870 and AD ~1930 with successively smaller glacier advances, but marked by distinct marginal moraines. This is in accordance with earlier studies from Folgefonna and also with old photos taken at the southern parts of Folgefonna. It seems like the southern part of Folgefonna had increased net mass-balance during the latest glacial expansion phase that culminated in AD 1940 (Tvede, 1972).

(8) The record of late Holocene glacial fluctuations may contribute to increased understanding of the coupling between oceanographic and atmospheric processes that led to the observed late Holocene decadal and millennial climate variability. Thus, it is apparent that high-resolution glacier reconstructions, especially from the last two millennia, should be adapted to a wider geographical area, involving glaciers in the range from continental to maritime climate regimes.

## Summary and conclusions

1. By using grain-size analysis and bulk density as proxies for former glacier size variations, it is shown that there is a potential for high-resolution glacier reconstructions in lakes where the LOI has its limitations (<~5%).

2. Sorting – mean anomalies have the prospective to track abrupt changes in the sedimentation environment of a lake and thereby validate the use of lake sediments to reconstruction of former glacier fluctuations.

3. By using a regression model between the average bulk values and a linear age-depth model based on the radiocarbon dates, it is possible to flux-correct the grain-size parameters, giving absolute

sedimentation rates. The approach also shows that the bulk density parameter is highly dependent upon sediment flux.

4. Basal radiocarbon dates from lake Dravladalsvatn indicate that glaciers were absent from the catchment shortly after 9600 cal. yr BP and that they reformed at 5200 cal. yr BP.

5. The early phase of mid-Holocene glacier growth was characterized by gradual glacier expansion leading to the first Subatlantic glacial event dated at 2300 cal. yr BP. This was a centennial-scale glacial readvance.

6. At 2200 cal. yr BP there is a significant change in glacier size, from a small glacier to glacier advances larger than at present. The record from the last 2200 years shows high-frequent glacial fluctuations at decadal and centennial time scales. It is indicated that the so-called "Medieval Warm Epoch" was a humid phase at northern Folgefonna, as glacier growth and decay during this time span was recorded. Altogether, the climate during the last 2200 years has been favourable for glacier growth at Folgefonna. The high-amplitude variation in ELA is therefore interpreted as a consequence of a more variable mode of the westerlies at the west coast of Norway.

7. The glacier net mass-balance for northern Folgefonna is modelled by using instrumental temperature and precipitation records from Bergen and Ullensvang back to AD 1800. Plotting the bulk density curve against the modelled glacier net mass-balance shows a remarkably similar pattern, where the maximum sediment yield was delayed with approximately 10 years due to a time lag between changes in glacier net-mass-balance and frontal response.

8. Dry bulk density (DBD) has the potential to resolve even small changes in mass-balance over short periods (sub centennial). Knowing that such rapid variations has limited effect on glacier size, increases the reliability of multi-decadal variations reconstructed from significant changes in sediment parameters.

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A black and white photograph of a mountain range. In the foreground, a large, textured glacier or snowfield slopes down from the right towards the center. The background shows a series of jagged mountain peaks under a bright, overcast sky. The overall scene is a high-altitude, glacial landscape.

## Paper 4

**Bakke, J., Dahl, S. O., Paasche, Ø.,  
Løvlie, R., and Nesje, A.** (submitted).

Late Weichselian and Holocene glacier  
fluctuations, equilibrium-line altitudes  
and palaeoclimate in Lyngen, northern  
Norway: a high-resolution composite  
chronology based on lacustrine and  
morpho-stratigraphical evidence.

*Quaternary Science Reviews*





# Late Weichselian and Holocene glacier fluctuations, equilibrium-line altitudes and palaeoclimate in Lyngen, northern Norway: a high-resolution composite chronology based on lacustrine and morpho-stratigraphical evidence

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Submitted to *Quaternary Science Reviews*

**Abstract:** Based on lacustrine and morpho-stratigraphical evidence from Lyngen in Troms, northern Norway, thirteen marginal moraines have been mapped and dated in front of Lenangsbreen in Strupskardet. Moraines M1-M3 and M8-M9 are suggested originally to be formed as ice-cored moraines, whereas M4-M7 and M10-M13 are suggested to be push and melt-out moraines. A chronological framework, taking into account a combination of relict periglacial features, former shorelines and related glacier-meltwater channels, lichenometry, and AMS radiocarbon dated lacustrine sediments spanning the last 20,000 cal. yr BP, has been established. The proglacial lake Aspvatnet was isolated from the sea c. 10,300 cal. yr BP, and the lacustrine sediments have been investigated by use of LOI at 550/950°C, magnetic susceptibility, water content, wet- and dry bulk density (DBD), and the magnetic parameters anhysteretic remanent magnetisation (ARM) and saturation remanent magnetisation (SIRM). There is in general good agreement between physical sediment parameters and magnetic parameters. DBD, a combination of medium and fine silt and the two statistical parameters 'sorting' and 'mean' have been used to construct a high-resolution glacier-fluctuation curve for the last 3800 cal. yr BP. Based on an accumulation-area ratio (AAR) of 0.6 and an adapted 'Little Ice Age ratio' (LR) approach, a continuous temperature-precipitation-wind equilibrium-line altitude (TPW-ELA) curve for the last 20,000 cal. yr BP have been constructed. Using an established exponential relationship between mean ablation-season temperature and mean annual solid precipitation at the ELA of Norwegian glaciers, variations in mean winter precipitation as snow are quantified using an independent proxy for summer temperature. Mean annual winter precipitation varied from 500 to 5000 mm water equivalents, and on average, Holocene estimates is c. 50% higher than similar figures from the Lateglacial. The two driest periods occurred during Heinrich events 1 (H1) (17,500-16,500) and 0 (H0) (13,000-12,200), whereas freshwater pulses to the North Atlantic had apparently no systematic impact on mean winter precipitation. Based on the winter precipitation curve from Lyngen, the atmospheric circulation responded to the SST lowering associated with H1 and H0 events in Lyngen with formation of talus-derived rock glaciers at sea level.

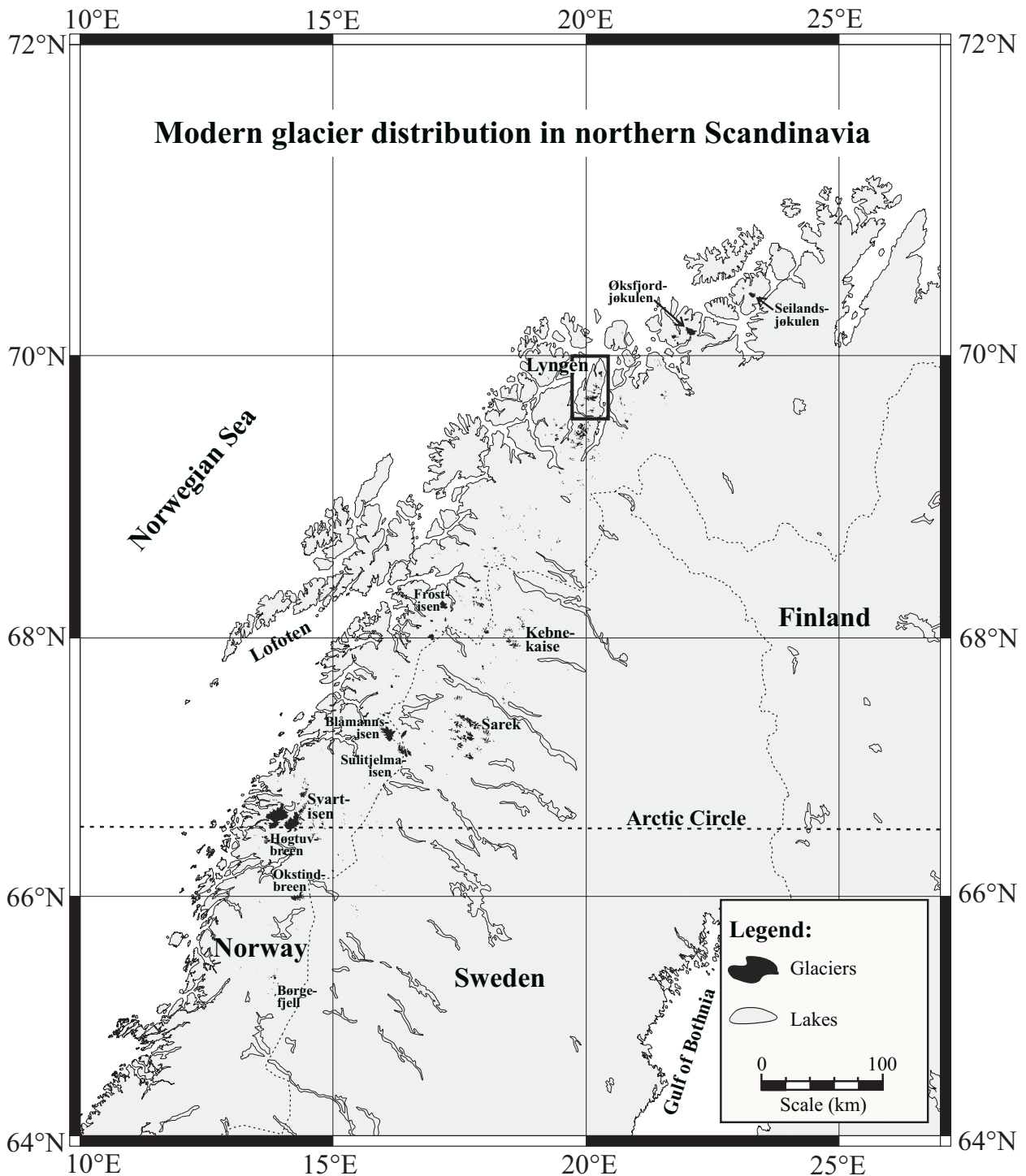
**Key words:** Lateglacial, Holocene, glacier fluctuations, equilibrium-line altitude, ELA, winter precipitation, summer temperature, NADW, Norway, northern Scandinavia, North Atlantic.

## Introduction

Along the coast of Troms, northern Norway (Fig. 1), the northward heat advection of air and water masses into the Norwegian Sea region produces some of the largest temperature anomalies in the world, with a mean January air temperature about 24°C warmer than the latitudinal mean in Tromsø (69° 40' N) (Encyclopaedia Britannica, 2003). The regional temperature and precipitation patterns are sensitive to variations of this heat advection. If it is reduced/turned off, the region is suggested to experience an 'instantaneous' shift from anomalously warm and open ocean climate conditions to a wide-spread cold climate regime with continuous sea-ice cover. Fluctuations in glacier size and equilibrium-line altitude (ELA) are primarily the

result of variations in mean ablation-season temperature and/or winter precipitation as snow. Because small glaciers have a short time delay on climatic changes (e.g. Nye, 1960; Paterson, 1994), fluctuations in glacier magnitude and ELA is suggested to be a sensitive proxy for variations in the atmosphere-ocean interaction in the Norwegian Sea region.

The Lateglacial and Holocene climate and glaciation history in northern Scandinavia has been studied by use of several approaches; reconstruction of ice sheets and deglaciation (Undås, 1938; Grønlie, 1940; Marthinussen, 1960, 1962; Holmes and Andersen, 1964; Andersen, 1968, 1975, 1979; Møller and Sollid, 1972; Sollid *et al.*, 1973; Rokoengen *et al.*, 1979; Vorren and Elvsborg, 1979; Corner, 1980; Andreassen *et al.*, 1985; Blake and Olsen, 1999;



**Figure 1** Map showing modern glacier distribution in northern Scandinavia. Bold frame shows the study area north of the fjord Kjosjen on the Lyngen Peninsula. The continental glaciers around Kebnekaise and Sarek in northern Sweden have the highest equilibrium-line altitudes (ELAs), whereas glaciers at Svartisen, Øksfjordjøkulen and Seilandsjøkulen have the lowest ELAs at present.

Fimreite et al., 2001; Plassen and Vorren, 2002; Vorren and Plassen, 2002), reconstruction of local glaciers (Karlén, 1973, 1976, 1981, 1988, 1997; Whalley, 1973; Karlén and Denton, 1975; Griffey, 1976; Worsley and Alexander, 1976; Corner, 1978; Griffey and Worsley, 1978; Gellatly et al., 1986, 1988, 1989; Gordon et al., 1987, 1996; Whalley et al., 1989, 1996; Ballantyne, 1990; Snowball, 1996; Snowball and Sandgren, 1996, 1997; Rea et al., 1998, 1999; Evans et al., 2002; Winkler et al., 2003), reconstruction of former sea-level changes (e.g. Marthinussen, 1960, 1962; Donner et

al., 1977; Hald and Vorren, 1983; Møller, 1984, 1985, 1986, 1987, 1989; Corner and Haugane, 1993; Kverndal and Sollid, 1993), reconstruction of summer temperature by biological proxies (e.g. pollen, macrofossils, diatoms, chironomids) (e.g. Vorren et al., 1988; Alm, 1993; Vorren and Alm, 1999; Seppä and Birks, 2001, 2002; Korhola et al., 2002; Seppä et al., 2002a, 2002b), reconstruction of former tree-line altitude by megafossils (Kullman, 1995, 1999, and references therein), oxygen-isotope records (Berglund et al., 1996; Shemesh et al., 2001;



Hammarlund *et al.*, 2002). Except for the 'Little Ice Age' (Gellatly, 1986; Whalley *et al.*, 1989, 1996; Ballantyne, 1990; Gordon *et al.*, 1996; Whalley and Kjølmoen, 2000) and Younger Dryas (Evans *et al.*, 2002), however, little is known about variations in the ELA of small glaciers during the Lateglacial and Holocene at the north-western coast of Scandinavia. With several small modern glaciers situated beyond the terminus of both the Skarpnes event and Younger Dryas Scandinavian ice sheets, the northern part of the Lyngen Peninsula (Figs. 1 and 2) in Troms is well suited for such an investigation.

The main objective of this study is to use lacustrine and morpho-stratigraphical evidence to obtain detailed information concerning the number, age and magnitude of Lateglacial and Holocene glacier fluctuations and ELAs in northern parts of the Lyngen Peninsula in Troms, northern Norway. Reconstructed ELAs combined with independent proxies for ablation-season temperatures will be used to quantify fluctuations in mean solid winter precipitation during the Lateglacial and Holocene (e.g. Dahl and Nesje, 1996). Variations in glacier magnitude, ELA and solid winter precipitation that are of regional importance for the atmosphere-ocean interaction in the North Atlantic region will be discussed.

## Study area

The study area is situated at the northern part of the Lyngen Peninsula in Troms, northern Norway (Figs. 1 and 2). About 141 km<sup>2</sup> of the Lyngen Peninsula is covered by glaciers at present (Østrem *et al.*, 1973), and increasing with altitude, numerous relict and active periglacial features are present in the area. The bedrock consists of a broad central belt of ophiolitic gabbro which is flanked on both sides by predominantly schistose metasediments (Randall, 1971; Chroston, 1972; Munday, 1974). The gabbro belt underlies the rugged mountains of Lyngsalpene ('the Lyngen Alps'), which are characterized by alpine peaks reaching altitudes up to 1800-1900 m.

Except for ice caps at the summits of Jiehkkevárri (altitude 1833 m) and Bälgesvárri (altitude 1625 m) south of the study area, cirque glaciers dominate the present glaciation on the Lyngen Peninsula. Large cirque glaciers like Sydbreen (c. 9 km<sup>2</sup>) and Strupbreen (c. 8.5 km<sup>2</sup>), however, may be regarded as valley glaciers. High-altitude glaciers are probable polythermal (subpolar), as they are thin and exist within the zone of discontinuous permafrost, whereas valley-outlet glaciers from the ice caps and low-altitude cirque glaciers, reaching down to 400 m, are inferred to be temperate.

Lake Aspvatn (altitude 35 m) is the second proglacial lake c. 6 km downstream of two partly coalescing cirque glaciers named Lenangsbreene (c. 2 km<sup>2</sup>) and three smaller cirque glaciers (c. 0.6 km<sup>2</sup>) in western parts of Strupskardet (Figs. 2-4).

Meltwater from these glaciers, including the dominant Lenangsbreene, drains through Blåvatn before it enters Aspvatn and then to the fjord Sør-Lyngan.

The climate in the study area is maritime at present, and an estimate based on the two meteorological stations Lyngseidet N (st. no. 9126) to the east and Tromsø (st. no. 9049) to the west, gives a mean summer temperature (Ts) (1 May-30 September) at sea level in Lyngen of c. 9.1°C (1961-1990) (DNMI, 1993b). Using an environmental lapse rate of 0.6°C/100 m (Sutherland, 1984), this gives a mean Ts at the present temperature-precipitation-wind ELA (TPW-ELA) (Dahl and Nesje, 1992) of 960 m at Lenangsbreene of c. 3.2°C (Fig. 5). Winter precipitation (Pw) (1 October-30 April) based on two meteorological stations in Ullsfjorden (station no. 9110, altitude 10 m; station no. 9111, altitude 6 m) and two at Lyngseidet (station no. 9125, altitude 6 m; station no. 9126, altitude 3 m) gives a mean (1961-1990) of 510 mm at sea level (DNMI, 1993a). Using a suggested mean exponential increase in winter precipitation with altitude of 8%/100 m in southern Norway (Haakensen, 1989; Dahl and Nesje, 1992), this gives a value of 1265 mm at the TPW-ELA of Lenangsbreene. However, as Lenangsbreene (N-NW) consist of two nearly coalesced cirque glaciers which receive a large additional amount of windblown snow, this underestimates the real value. The temperature is regionally more consistent than precipitation, and by using a mean ablation-season temperature of 3.2°C at the TPW-ELA of Lenangsbreene in equation 1 (see below), this corresponds to a winter precipitation (Pw) of c. 2260 mm.

Based on mass-balance measurements performed in 1998 (Kjølmoen, 2000), the corresponding TPW-ELA at Strupbreen (NE) and Koppangsbreen (E-SE) are 760 and 740 m, respectively (Fig. 5), or about 200 m lower than that of Lenangsbreene. The effect of solar radiation on north-facing compared to south-facing cirque glaciers is calculated to about 250 m (or 1-1.5°C) in Troms (Andersen, 1968). However, this effect is suggested to be negligible among the 3 investigated glaciers, and the modern ELAs of Lenangsbreen, Strupbreen and Koppangsbreen are suggested to reflect a prevailing snow-bearing wind direction from the southwest.

## Methods

The reconstruction of Lateglacial and early-Holocene glacier fluctuations and winter precipitation at Strupskardet involved several approaches:

- Air photographs (Widerøe, 1962) and field observations were combined to produce a detailed glacial-geomorphological map in scale 1:50,000 for the Strupskardet catchment with emphasis on former marginal moraines, glacier-meltwater channels, glaciofluvial deposits and landforms (deltas/



**Figure 2** Map showing the northern part of the Lyngen Peninsula. Dotted lines show the suggested ice margins formed by the Scandinavian Ice Sheet during the Skarpmes event and solid lines indicate the ice margin(s) during the Tromsø-Lyngen event (Younger Dryas). Black dots indicate sites used in the equidistant shore-line (ESL) diagram (Figure 7). The studied area in Strupskardet is located well beyond the Skarpmes event ice margin.

sandurs), raised marine terraces, beach ridges and shorelines, rock glaciers and tundra polygons.  
- The timing of Lateglacial events linked to sea level was sorted out based on the regional sea-level history

recorded around Arnøy (Kverndal and Sollid, 1993) and from Lyngen (Corner and Haugane, 1993) and correlated with observations in this study from Russelv to Skardmunken (Fig. 2).





**Figure 3** Photograph showing outer parts of Strupskardet and Aspvatnet from above. The ice-cored moraine M3 is prominent at the left side of the valley, whereas the right side of the valley is dominated by talus fans and talus-derived rock glaciers. The blue colour in Aspvatnet is caused by input of glacier-derived meltwater. Photo: Jostein Bakke.

- Lichenometry based on *Rhizocarpon geographicum* (e.g. Ballantyne, 1990; Matthews, 1994) was used to establish relative-age chronologies for the marginal moraines in front of Lenangsbreen with particular emphasis of identifying moraines older than the 'Little Ice Age' maximum in Lyngen.

- The deposition of distinct marginal moraines is related to (rather) short periods when the glacier is in steady-state, whereas younger glacier (e.g. the 'Little Ice Age') advances may have erased older marginal moraines (e.g. Matthews, 1991; Dahl and Nesje, 1994). Hence, proglacial sites (lacustrine and terrestrial) beyond the maximum extent of the suggested glacier advances within the studied time span are taken into account to obtain continuous records of glacier fluctuations. The methods related to proglacial sites are based on a conceptual model of glacier-meltwater induced sedimentation in which the minerogenic (nonorganic) component of the sediments is related to the presence of a glacier in the catchment (Karlén, 1981; Leonard, 1985; Dahl et al., 2003).

- Two cores were retrieved from Aspvatnet (Fig. 3) during the late summer of 2000 by use of a modified piston corer (Nesje, 1992) taking up to 6 m long tubes with diameter 11 cm. The two cores were 4.3 and 4.5 m long, respectively, and were taken from the deepest part of the lake. Both cores were subsampled

at 0.5 cm intervals using volume determinant sampling equipment in the laboratory for a number of analyses:

- For loss-on-ignition (LOI) (Dean, 1974) the samples were ignited for 1 hour at 550°C.

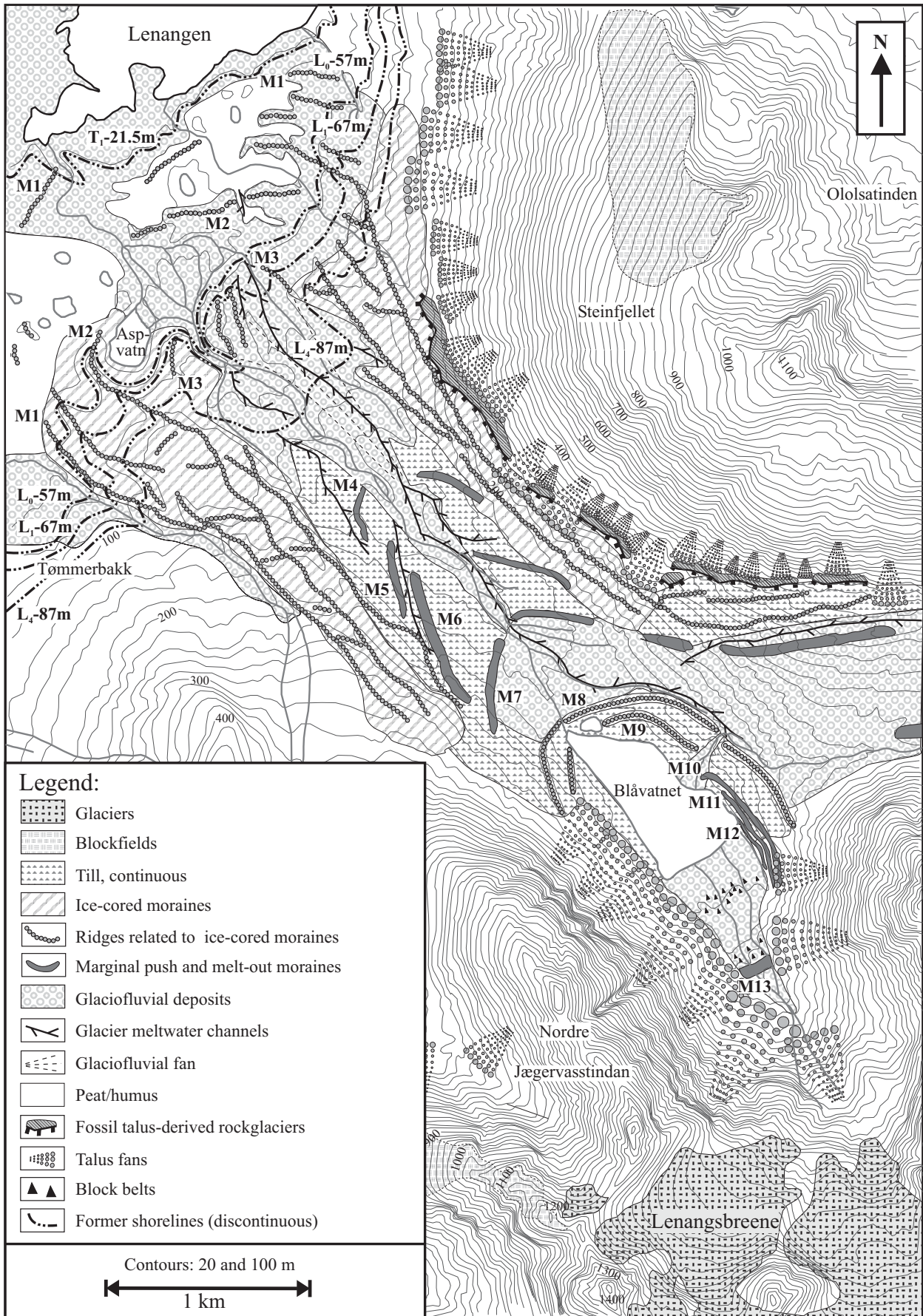
- Wet and dry bulk density (DBD) was measured before and after drying.

- Water content was calculated as per cent of dry sample weight.

- Fine-grained sediments were analysed using a 'Micromeritics SediGraph 5100 Particle Size Analyze System' (MasterTech, 1993). This system measures particle diameter based on Stoke's law in a range from 300 to 0.1 micrometer ( $\mu\text{m}$ ) equivalent spherical diameters by shooting x-rays through suspended material.

- Prior to sub-sampling, surface magnetic susceptibility (MS) was determined at 5 mm intervals with an in-house constructed automatic instrument using a Bartington MS2E sensor on a cleaned surface (2/3 of core) covered with a thin plastic film (<0.1 mm thick). A total of 97 cubic plastic boxes (2x2x1.8 cm) were retrieved. Sampling intervals varied from 2 cm (0-100 cm and 210-270 cm) and between 10 and 20 cm for the remaining core. Sample weights were converted to dry weight using the average water content value for parallel samples, whereas MS was determined on a KLY-2 induction balance (sensitivity





**Figure 4** Glacial-geomorphological map showing the valley Strupskardet and the two proglacial lakes Blåvatnet and Aspvatnet. The valley floor is dominated by glaciofluvial deposits and till. The marginal moraines M1-M3 and M8-M9 are mapped as ice-cored moraines indicating continuous permafrost during the time of deposition. The moraine M13 is the only moraine found from the late-Holocene glacial advances, and was most probably deposited during the 'Little Ice Age' maximum.



$4 \times 10^{-8}$  SI). Anhyseretic remanent magnetisation (ARM) was acquired in a 0.1mT DC-field and 100 mT AC-field using a 2G af demagnetiser. Saturation isothermal remanent magnetisation (SIRM) was imposed in a 3T Redcliffe pulse magnetiser. Backfield IRM was imposed antiparallel to SIRM in a 100mT field using a solenoid and in  $-300$ mT field using the pulse magnetiser. The magnitude of these parameters is a function of the type(s) and concentration of the magnetic minerals. The concentration independent ratios ARM/SIRM (magnetite grain size),  $S-0.3T$  (IRM-300mT/SIRM) and  $S-0.1T$  (IRM-100mT/SIRM) have been calculated. Thermomagnetic curves were run on 7 dried bulk samples in a horizontal Curie balance, whereas heating/cooling were done in air with heating rate  $15^\circ\text{C}/\text{min}$  and magnetic fields between 450 to 600 mT.

- Glacier ELAs are reconstructed using an accumulation-area ratio (AAR) of 0.6 (Dahl *et al.*, 1997). To reconstruct former ELAs of debris-mantled glaciers in high-mountain regions may be difficult (Clark *et al.*, 1994; Benn and Lehmkuhl, 2000). Because of rather well-defined landforms, however, remnants of ice-cored marginal moraines were used to reconstruct former ELAs in the study area. The calculation of the glacier area distribution was carried out using the vector-based GIS program MapInfo 6.0 on an N-50 map datum.

- Winter precipitation estimates were calculated based a close exponential relationship between mean ablation-season temperature  $t$  ( $^\circ\text{C}$ ) (1 May–30 September) and mean solid winter precipitation  $A$  (m water equivalent) (1 October–30 April) at the ELA of Norwegian glaciers in maritime to continental climatic regimes (both temperate and polythermal glaciers) (Liestøl in Sissons, 1979; Ballantyne, 1989; Dahl *et al.*, 1997, and references therein):

$$A = 0.915e^{0.339t} \quad (r^2 = 0.989, P < 0.0001) \quad (\text{Eq.1})$$

Eq. 1 (the 'Liestøl equation') implies that if either the winter precipitation or the ablation-season temperature at the ELA is known, the unknown variable can be calculated. It also implies that if the former ELA is known, it is possible to quantify how the winter precipitation has fluctuated if an independent proxy for mean ablation-season temperature is used in the calculation (Dahl and Nesje, 1996; Dahl *et al.*, 1997).

## Results

The valley of Strupskardet and the neighbouring mountains contain a number of different active, inactive, and relict landforms that are related to climate change (Figs. 3 and 4). The mountains surrounding Strupskardet are rather flat and undulating with autochthonous blockfields on the

summits and allochthonous blockfields at the steeper upper slopes. Along the northern valley face of Strupskardet and at both sides of Blåvatn there are active talus formation and several snow avalanche tracks. There is no active moraine formation in front of Lenangsbreen at present. Along the valley floor of Strupskardet, however, the main features are thirteen marginal moraines (M1-M13).

## Moraine chronology

### Moraines M1-M3

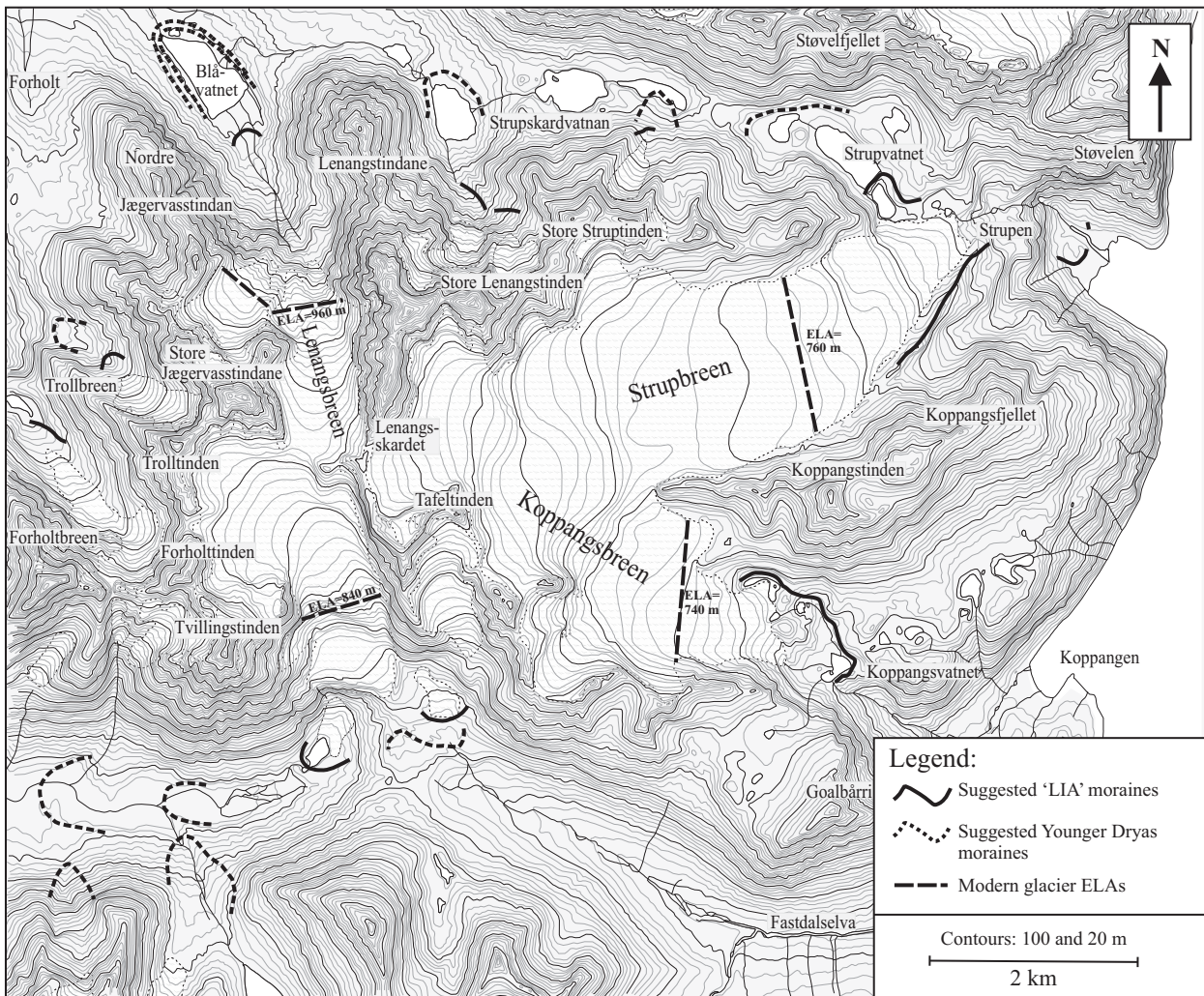
In the outer part of Strupskardet, three more-or-less continuous ridges named M1-M3 can be traced from an altitude of 395 to 20 m along the southern valley side (Fig. 4). In between the parallel main ridges, several transverse ridges are present in the southern part of the complex. The ridges are located close to each other in the upper end. However, the distance increases in the flat areas closer to the fjord. The three ridges can also be followed along the northern valley side, but are less distinct because of talus formation and/or features suggested to be relict rock glaciers (see below). The three ridges have a mean gradient of 40 m/km in the upper end. On both sides of Strupskardet a terrace, suggested to be the result of marine abrasion (see below), has been cut into the ridges c. 90 m a.s.l. Here, the angle of the ridge complex changes, and a chaotic and rather flat area characterised by a number of depressions, inferred to be the result of melt-out of buried ice, is present below this altitude.

The three ridges M1-M3 are suggested to be ice-cored moraines formed by polythermal (subpolar) valley glaciers. Ice-cored moraine systems are found in the Arctic and at high-alpine sites worldwide (Østrem, 1964; Boulton, 1967; Dowdeswell and Sharp, 1986; Hambrey *et al.*, 1999). The formation and preservation of ice-cored moraines are commonly related to cold-based glacier margins found in permafrost regions (Østrem, 1964). The debris accumulation is related to supraglacial sedimentation or melt-out of englacial debris.

### Moraines M4-M7

The M4 ridge is mapped at a lowest altitude of 120 m. The ridge is between 2-4 m high, and with the steepest side dipping  $40-30^\circ$  towards the fjord. Only separated by the modern meltwater stream, the ridge can be traced on both sides of the valley. Dry channels inferred to be glacier meltwater streams at the distal side of the ridge disappear at an altitude of c. 87 m. Ridges M5-M7 have a similar asymmetrical appearance, size, and occurrence at the valley floor as M4, but lie on top of a glaciofluvial fan with the surface dipping towards the fjord at an angle of  $3-5^\circ$  (Fig. 4).

These ridges are all inferred to be push- and melt-out moraines formed by advancing glaciers. The suggested meltwater channel distal to the M4 moraine is linked to the M4 glacier.



**Figure 5** Map showing the modern TPW-ELAs and reconstructed ice-marginal moraines from the 'Little Ice Age' and Younger Dryas at the ice field consisting of Strupbreen, Lenangsbreen, Koppangsbreen and several smaller glaciers. The TPW-ELA is about 200 m lower at the eastern side of the Lyngen Peninsula as a result of the prevailing snow-bearing wind direction from southwest with enhanced leeward accumulation of snow.

### Moraines M8-M9

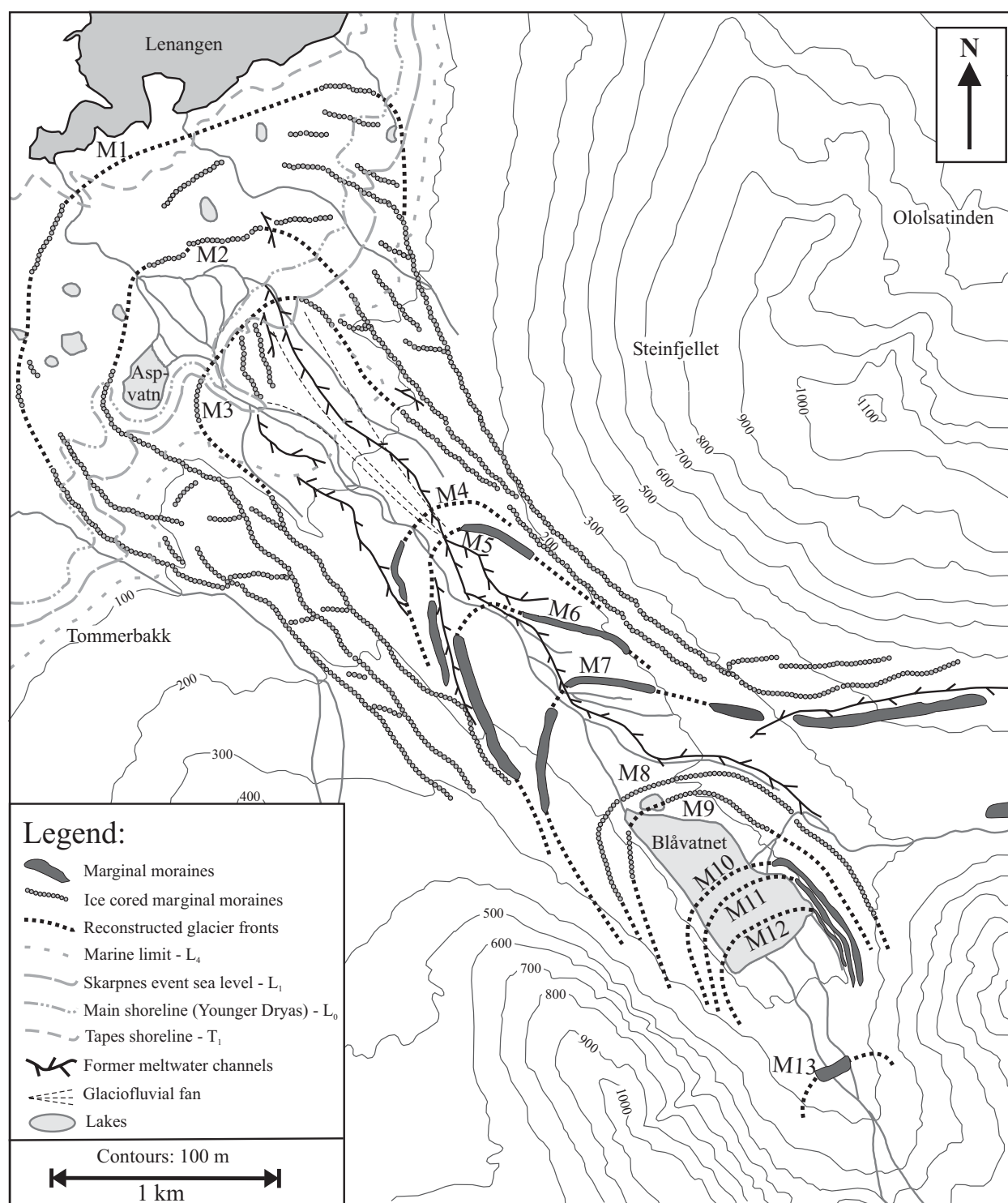
Separated by c. 50 m, the very distinct ridges M8 and M9 are curved around Blåvatnet (Fig. 4). Both ridges consist of angular boulders varying in size from 1-15 m<sup>3</sup>, whereas fine-grained sediments apparently are lacking. The present meltwater stream from Blåvatnet partly drains through M8 and M9. M8 can be traced to a maximum altitude of 290 m along the northeastern valley side. Between M8 and M9 there are melt-out features and stone circles suggested to be relict frost polygons (see below). At present, fine material is abundant in Blåvatnet and along the valley sides inside M9. A channel inferred to be a former meltwater stream from upper Strupskardet can be followed between M8 and the north-eastern valley side. Further downstream this former meltwater channel has been mapped north of the present meltwater stream in Strupskardet to an altitude of 57 m, where it disappears. Lichen diameters on M8 and M9 are larger than 250 mm.

Both M8 and M9 are interpreted to be marginal moraines. Lack of fine material from Blåvatnet in these moraines may indicate supra or englacial

transport and deposition of the angular boulders (e.g. Lowe and Walker, 1997). Because of the large amounts of debris, M8 and M9 may be classified as 'latero-frontal dump moraines' formed as supraglacial debris slid, rolled or flowed towards the glacier margins (Boulton and Eyles, 1979). Supraglacial deposition and lack of fine material in M8 and M9 may both indicate that a rock avalanche took place on the surface of a polythermal (subpolar) glacier with subfreezing conditions in the ablation zone. This is supported by relict frost polygons between the ridges (see below), and M8 and M9 may be regarded as ice-cored moraines. By following the lateral moraines at the northern side of Blåvatnet, a rather steep glacier profile has been reconstructed. This may indicate polythermal conditions, as a steep gradient is more likely if the glacier sole near the glacier front is frozen to the substratum.

The relict meltwater channel between M8 and the northern valley side links M8 and M9 to a former sea level of 57 m, whereas melt-out features between the ridges indicate a minor retreat in between deposition of the two ridges (Fig. 4).





**Figure 6** Map showing the reconstructed glaciers at thirteen glacier stages obtained from the mapped moraine chronology in Strupskardet. Former sea levels during the Lateglacial and Holocene are indicated with dotted lines in the lower part of the valley towards Lenangen. The former sea level at 87 m a.s.l. has eroded the outermost ice-cored moraines M1-M3.

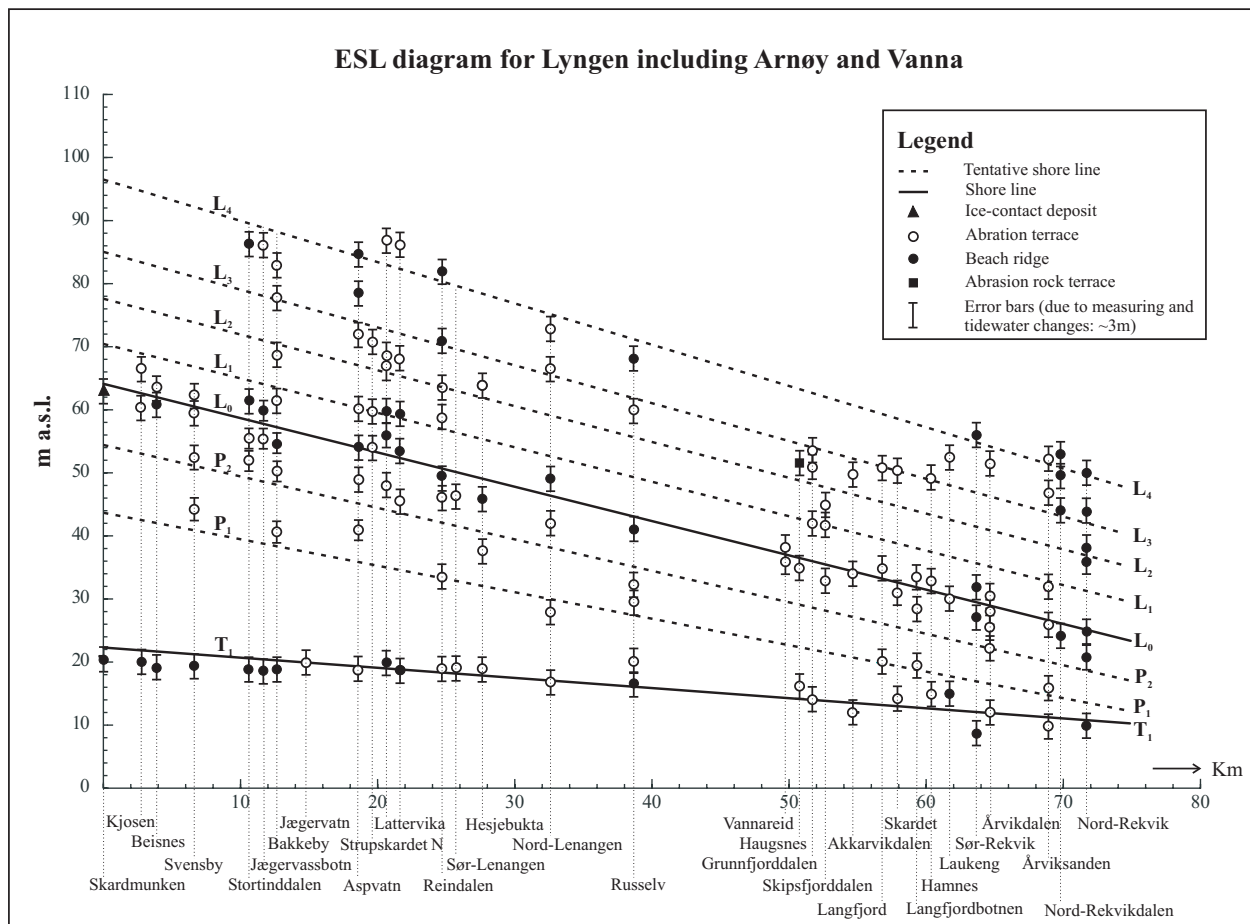
#### Moraines M10-M12

The ridges M10, M11 and M12 are located more or less continuously along the northeastern shore of Blåvatn (Fig. 4), whereas they are missing along the talus-covered southwestern shore. M10 is the most distinct ridge and can be traced from 30 to 3 m above Blåvatn for about 200 m. M11 and M12 are somewhat shorter, but like M10, they vary in height from 0.5–2.0 m and consist of unsorted sediments. The distance between the ridges is about 50 m, and the

lower part of M12 disappears into Blåvatn c. 40 m beyond the modern glaciofluvial delta in the upper end of the lake. Lichen diameters on the three ridges are larger than 200 mm.

Based on size and location along the northeastern shore of Blåvatn, M10, M11 and M12 are inferred to be push and melt-out marginal moraines formed by tempered glaciers (Fig. 4). In contrast to the moraine ridges M8 and M9, M10-M12 indicates low-gradient glaciers calving into Blåvatn.





**Figure 7** Equidistant shore-line (ESL) diagram for Lyngen, Arnøy and Vanna. The measurements from Russelv to Skardmunken are from this study (82 measurements), whereas measurements from Arnøy and Vanna are adopted from Corner and Haugane (1993) and Kverndal and Sollid (1993). The most prominent former sea levels are the L<sub>0</sub> line/Main shoreline from the Younger Dryas and T<sub>1</sub> line from the Tapes transgression.

### Moraine M13

Ridge M13 mainly consists of angular blocks varying from 0.5-10 m in diameter, whereas fine material is sparse. The ridge is located about 600 m upstream of Blåvatn (Fig. 4), and has a maximum height of 15 m. The surface of M13 is rather chaotic, and there is no obvious difference in slope angle on the proximal and distal parts of the ridge. Talus dominates on both valley sides of M13. Maximum observed lichen diameter on the ridge is 38 mm (see below). Two block belts are present between M13 and Blåvatnet.

M13 is inferred to be a push moraine formed by an advancing glacier.

### Reconstructed glacier events

Based on the thirteen marginal moraines, reconstructed glacier events from Strupskardet are shown in Figure 6. They can be divided into 4 main stages:

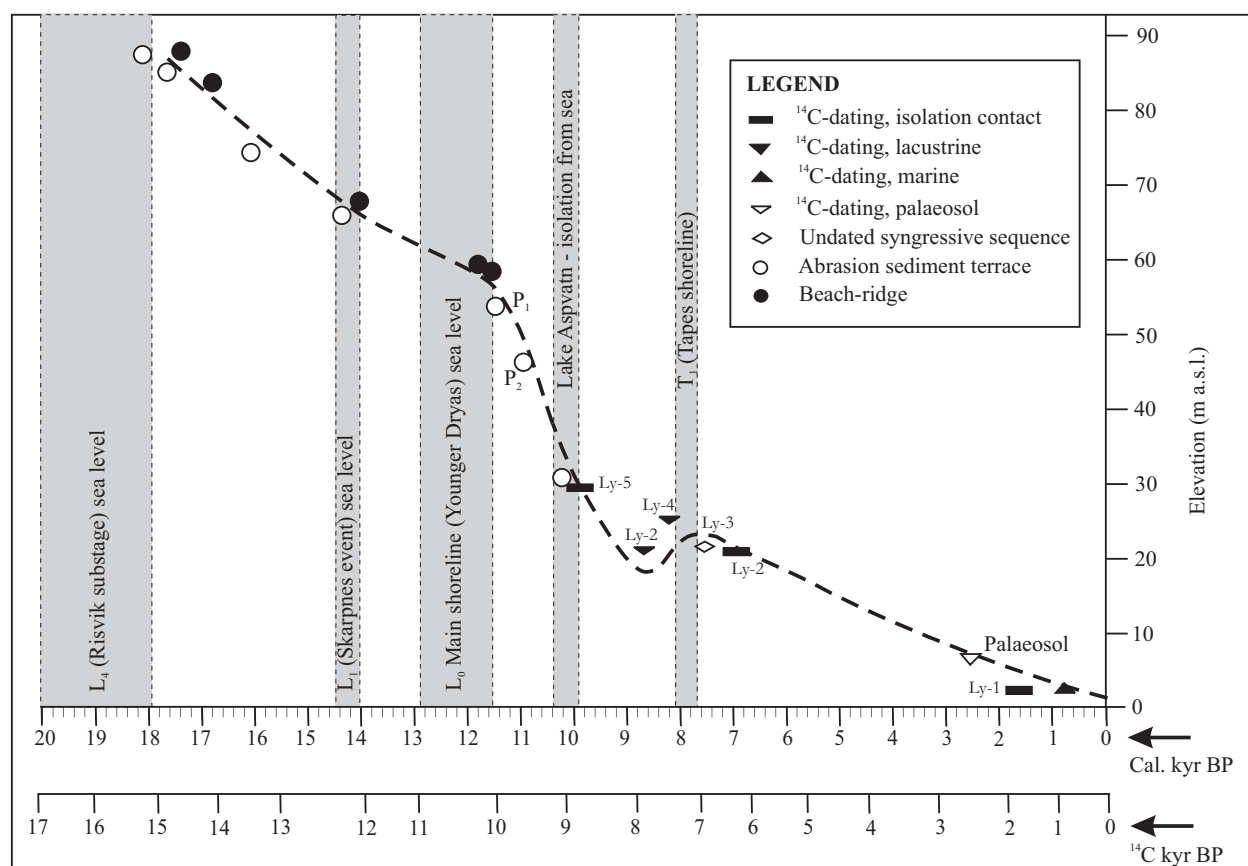
- M1 to M3 are ice-cored moraines.
- M4 to M7 are push and melt-out moraines.
- M8 to M9 are likely ice-cored moraines.
- M10 to M13 are push and/or melt-out moraines.

The reconstructions shown in Figure 6 make the basis for calculation of former ELAs adjusted for land uplift.

### Lateglacial and early-Holocene sea-level history in Strupskardet

The pattern of relative sea-level changes in northern Norway is generally known based on a number of shoreline diagrams (Marthinussen, 1960, 1962; Andersen, 1968, 1975; Møller and Sollid, 1972; Sollid et al., 1973; Møller, 1985, 1987; Corner and Haugane, 1993; Kverndal and Sollid, 1993) and shoreline displacement curves (Marthinussen, 1962; Donner et al., 1977; Corner, 1980; Hald and Vorren, 1983; Møller, 1984, 1986, 1987; Vorren and Moe, 1986). In northern Troms, three Holocene shoreline displacement curves have been constructed (Corner, 1980; Hald and Vorren, 1983; Corner and Haugane, 1993). As many areas in northern Norway lack direct dating evidence for Lateglacial sea-level changes, several reconstructions are related to the early works of Marthinussen (1960, 1962).

From Arnøy (Fig. 2), Kverndal and Sollid (1993) constructed a Late Weichselian deglaciation model based on the altitudinal distribution of blockfields, weathered surfaces, glacial striae, erratics and an equidistant shoreline (ESL) diagram. Arnøy is located 20 km north of the Lyngen Peninsula, and by extrapolating the ESL diagram southwards to



**Figure 8** Shoreline-displacement curve from Strupskardet modified from Corner and Haugane (1993). The Holocene part of the sea-level curve is based on coring of isolation basins, whereas the Lateglacial sea-level curve is constructed based on the ESL diagram in Figure 7. Radiocarbon dates of shells from core 1 in Aspvatnet are in agreement with the proposed early-Holocene sea-level history by Corner and Haugane (1993).

Ullsfjorden along the observed land-uplift gradient, it was possible to use the same age estimates for Lateglacial sea levels in this area.

From the Lyngen Peninsula, new measurements at 17 localities from Russelv to Skardmunken (Figs. 2 and 7) have been put into an extrapolated ESL diagram from Arnøy. Raised abraded sediment terraces and beach ridges are the most prominent features, but also abraded rock terraces and ice-contact deposits contribute with information regarding Lateglacial sea-level changes. The two lines  $L_0$  and  $T_1$  are distinct shorelines, whereas the lines  $L_1$ ,  $L_2$ ,  $L_3$ ,  $L_4$ ,  $P_1$  and  $P_2$  are tentative lines based on Marthinussen (1960) and Kverndal and Sollid (1993).

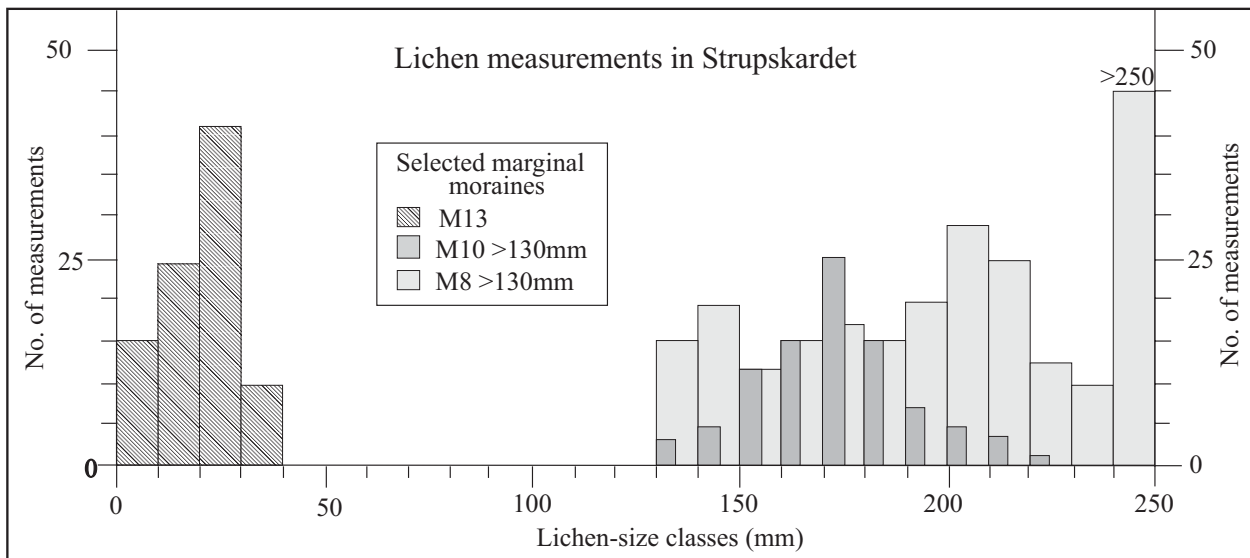
Holocene sea-level changes are based on the shoreline displacement curve constructed by Corner and Haugane (1993) from sites close to the study area in Strupskardet. As shown in Figure 6, the  $L_0$  line/Main shoreline corresponds to the Younger Dryas sea level,  $T_1$  to the Tapes line, whereas  $L_1 - L_4$  are older than the  $L_0$  line.  $P_1$  and  $P_2$  are lines representing Preboreal sea levels. The  $L_4$  line on Arnøy has been extrapolated inland to Jægervassbotn (Figs. 2 and 7), and is correlated to the Risvik substage with an estimated age of c. 20,000–18,000 cal. yr BP in Porsangerfjorden in Finnmark (Sollid et al., 1973; Andersen, 1979). At Svensby (Figs. 2 and 7), the  $L_1$  line is suggested to represent the sea level during

the Skarpnes event (c. 14,500–14,000 cal. yr BP) (e.g. Vorren and Plassen, 2002). Between the  $L_4$  and  $L_0$  lines in Strupskardet, however, no marked shorelines have been found, and the glacier in Ullsfjorden is therefore suggested to have retreated without any major halts or readvances at least to Svensby or Skardmunken (Paasche et al., submitted). As a consequence, the outer parts of Strupskardet may have been ice-free during at least the last 20,000 cal. yr BP.

#### Raised shorelines in Strupskardet

The presence of beach ridges, melt-out features, abraded terraces and glaciofluvial deltas show that the outer parts of Strupskardet have been modified by Lateglacial and Holocene sea-level changes (Fig. 4). The highest former shoreline is marked by beach ridges and abraded terraces at an altitude of 91 m, whereas a distinct former sea level is marked by beach ridges at an altitude of 87 m. The latter terrace is especially prominent at the northern flank of the valley. Based on the extrapolated ESL diagram, this high sea level may be connected to the  $L_4$  line from the Risvik substage (Figs. 7 and 8).

A minor halt in the sea-level lowering is seen as abraded terraces at 80 m a.s.l., whereas a distinct former sea level at an altitude of 69 m is evident from beach ridges, abraded terraces and melt-out forms in the ice-cored moraine complex (M1–M3) (Fig. 4). Based on the extrapolated ESL diagram, this sea level



**Figure 9** Lichen-size measurements on *Rhizocarpon geographicum* at selected moraines in Strupskardet. The M13 moraine contains lichens with maximum diameter of 40 mm, indicating an age close to the 'Little Ice Age' glacier maximum in Lyngen (e.g. Ballantyne, 1990). Similar measurements on the M10 and M8 moraines yield maximum lichen diameters >230 and >250 mm, respectively, that are well beyond the reliable range of modern growth curves in Norway (e.g. Innes, 1985; Matthews, 1994; Winkler et al., 2003).

corresponds to the  $L_1$  line from the Skarpsnes event (Figs. 7 and 8).

The most prominent former sea level is found at an altitude of 57 m, and represents the  $L_0$  line/Main shoreline from the Younger Dryas (c. 12,800-11,500 cal. yr BP) (Figs. 7 and 8). This former sea level is marked several places on the northern part of the Lyngen peninsula with up to 1.5 m high beach ridges, large abraded terraces and several glaciofluvial deltas. At the southern slope of Aspvatnet (Fig. 4), some less distinct abraded terraces are found at altitudes of 52 and 48 m, and probably represent the two Preboreal sea levels  $P_1$  and  $P_2$  (Figs. 7 and 8). The next distinct former sea level, at an altitude of 22 m, is represented by abraded terraces and a change in the angle of deposition in two alluvial fans built out by rivers from Strupskardet. This prominent sea level is suggested to represent the  $T_1$  line from the Tapes transgression (c. 8,000 cal. yr BP) (Corner and Haugane, 1993) (Fig. 8).

The  $L_0$  line/Main shoreline and later sea-level history are in accordance with the record outlined by Corner and Haugane (1993) from the Strupskardet area.

## Distribution of relict periglacial features

Several relict features related to a periglacial environment are recorded in and around Strupskardet. The vertical and horizontal extent of these features may give important independent information regarding the climate and timing when the marginal moraines in Strupskardet were formed.

### Rock glaciers

Along the south-southwest facing valley side of Strupskardet, a series of 12 'hanging' lobes, mainly consisting of blocks and boulders with distinct fronts

towards the valley, can be mapped for about 3 km (Fig. 4). The front of individual lobes are up to 10-30 m high with slope angles varying from 30-45°, and with a plateau behind consisting of transverse ridges with up to 5 m deep depressions in between. The distance from the back wall to the front of the lobe plateaux vary from 20 to 70 m, and the altitude of individual lobes drops from about 550 m in inner parts to 140 m towards the fjord. More-or-less active talus fans are present in the valley slopes uphill from the lobes, and several places these fans have filled in the lobe plateaux. In one case, a rather small lobe is found below another, but all described lobes are situated beyond the M1 moraine (Fig. 4).

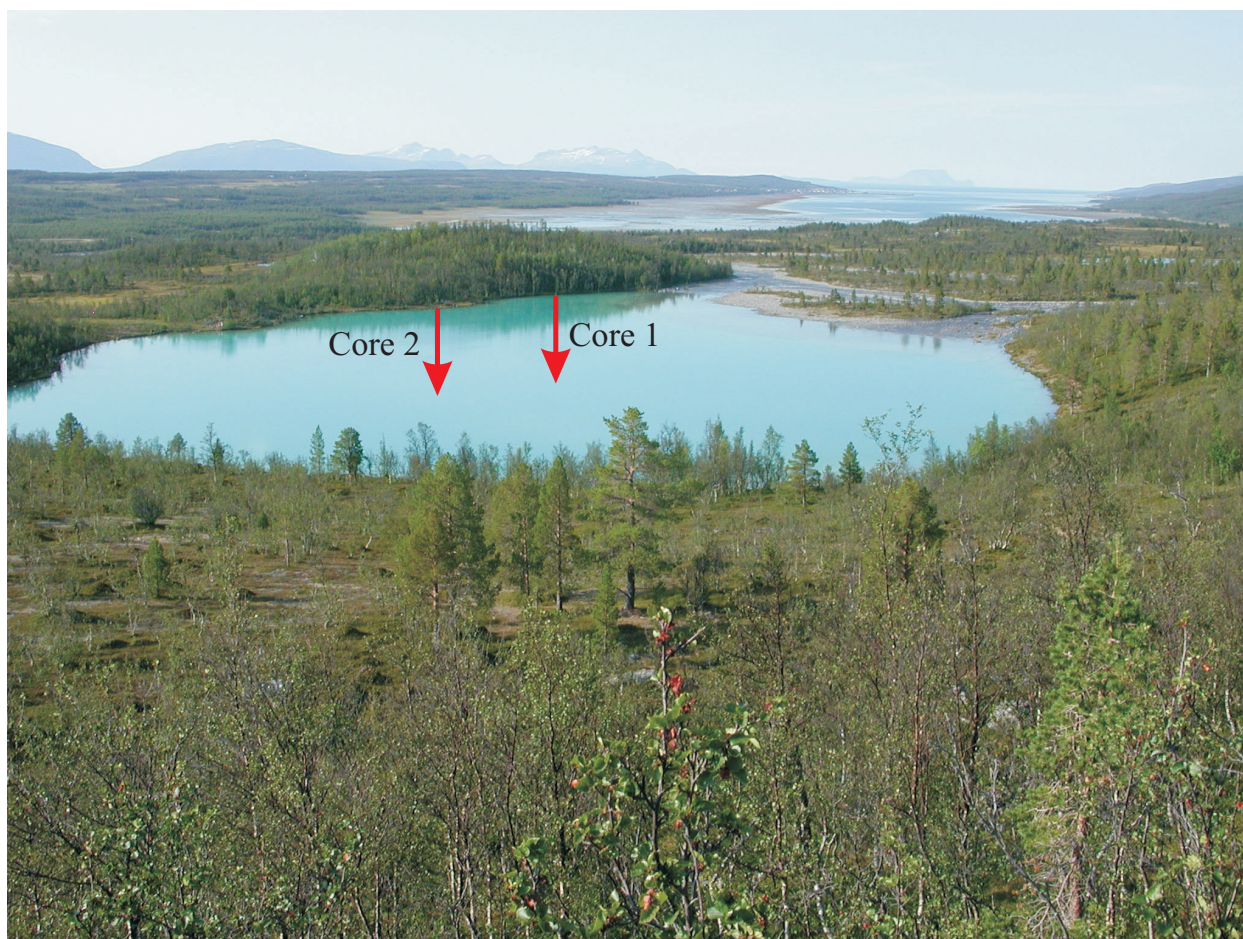
The described lobes are suggested to be relict talus-derived rock glaciers. Their initiation was dependent on sufficient material supply from the south-southwest facing back wall, and developed in a continuous permafrost environment (Fig. 4). The altitude of individual rock glaciers coincides with the vertical extent of the M1 glacier, implying that the rock glaciers may have been formed in association to the glacier deposition this moraine.

### Frost polygons

Circles of sorted stones are interpreted to be periglacial frost polygons and are, except for one site at/in Aspvatn (altitude 35 m), commonly found in Strupskardet between the  $L_0$  line/Main shoreline (altitude 57 m) and beyond the M8 moraine at the outlet of Blåvatn (Fig. 4). The polygons appear in glaciofluvial sediments with individual diameters varying from 0.2 to 2 m.

Frost polygons are generally used as permafrost indicators, and occurrence of these features are previously used in Troms and Finnmark as an indicator of ice-free conditions during the Younger Dryas (e.g. Kverndal and Sollid, 1993). Based on their geographical distribution in Strupskardet, the frost





**Figure 10** Aspvatnet with the inlet and outlet in the northern part of the lake. The fjord Lenangen in the background. Red arrows indicate positions of the two cores used in this study. Photo: Jostein Bakke.

polygons are suggested to be predominantly relict.

The anomalous frost-polygon site at shallow depth near the inlet in Aspvatn is normally covered by the lake in summertime during high runoff from the glaciers, whereas the site is above the water level during the winter season. Hence, during the summer season frost-sensitive silt is deposited at the site, whereas during cold and dry winters this silt is exposed to frost sorting even at present. The up-frozen coarser material at the Aspvatn site originates from older (glacio-) fluvial sediments.

## Lichenometry

To distinguish marginal moraines younger and older than the 'Little Ice Age', systematic measurements of largest lichen diameters of *Rhizocarpon geographicum* according to the methods described by Matthews (1994) was carried out on the moraines M8, M10 and M13 (Fig. 9). In addition, test measurements were carried out on M11 and M12. On M8 the largest diameters were >250 mm, on M10 >200 mm, and on M13 the maximum measured diameter was 38 mm. The observations from M11 and M12 were comparable to M10.

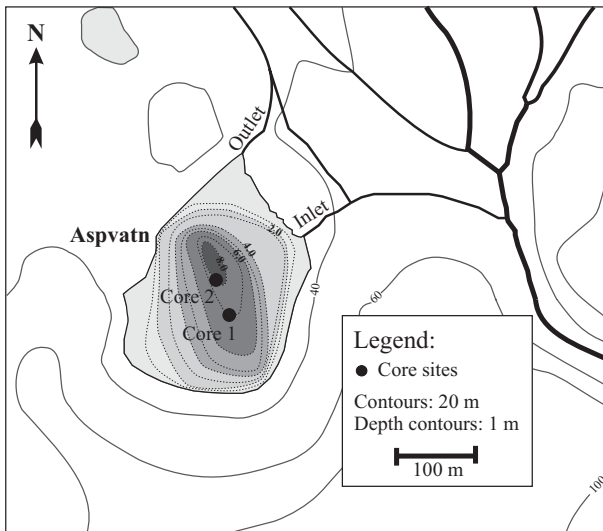
The maximum diameter of 38 mm on M13 fits with the inferred 'Little Ice Age' maximum between

A.D.1890 -1928 (e.g. Ballantyne, 1990), whereas the lichen diameters of M8 to M12 are well beyond the reliable range of modern growth curves in Norway (e.g. Innes, 1985; Matthews, 1994; Winkler *et al.*, 2003).

## Lacustrine sediments from lake Aspvatnet

Aspvatnet is an almost circular lake with diameter c. 250 m inferred to be a former kettle hole formed between the ice-cored moraines M2 and M3 (Figs. 3, 4 and 10). The lake is surrounded by gentle slopes with angles of 3-7°. The inflow and outflow are in the northern end of the lake, and it has a maximum water depth of 12.3 m (Fig. 11).

In addition to Aspvatnet, the Strupskardet catchment includes five lakes with input of glacier meltwater. Among these, Blåvatnet is of greatest importance as it acts as a sediment trap for the glacier-meltwater streams in the catchment before entering Aspvatnet. Hence, Aspvatnet is a suitable lake basin to study glacier-meltwater induced sedimentation from suspension (e.g. paper 1). Because of their size, Lenangsbreene are suggested to dominate the input of glacier-meltwater induced sediments in Aspvatnet (Fig. 4). Whenever Lenangsbreene advances across Blåvatnet, however, the sedimentation in Aspvatnet



**Figure 11** Bathymetric map of Aspvatnet. Thick lines show the main stream channels on the fan delta. Both the inlet and outlet in Aspvatnet are in the northern part of the lake.

will be dominated by coarser clastic sediments.

A distinct terrace 57 m a.s.l. around Aspvatnet is inferred to result from marine abrasion in the same period as a coarse fan-shaped delta along the present meltwater stream from Strupskardet was formed at this altitude. The fan-shaped delta was formed when the meltwater stream cut through M3, and the resulting gully appears to have been stable since because the modern stream lack competence to transport blocks up to 5 m in diameter present on

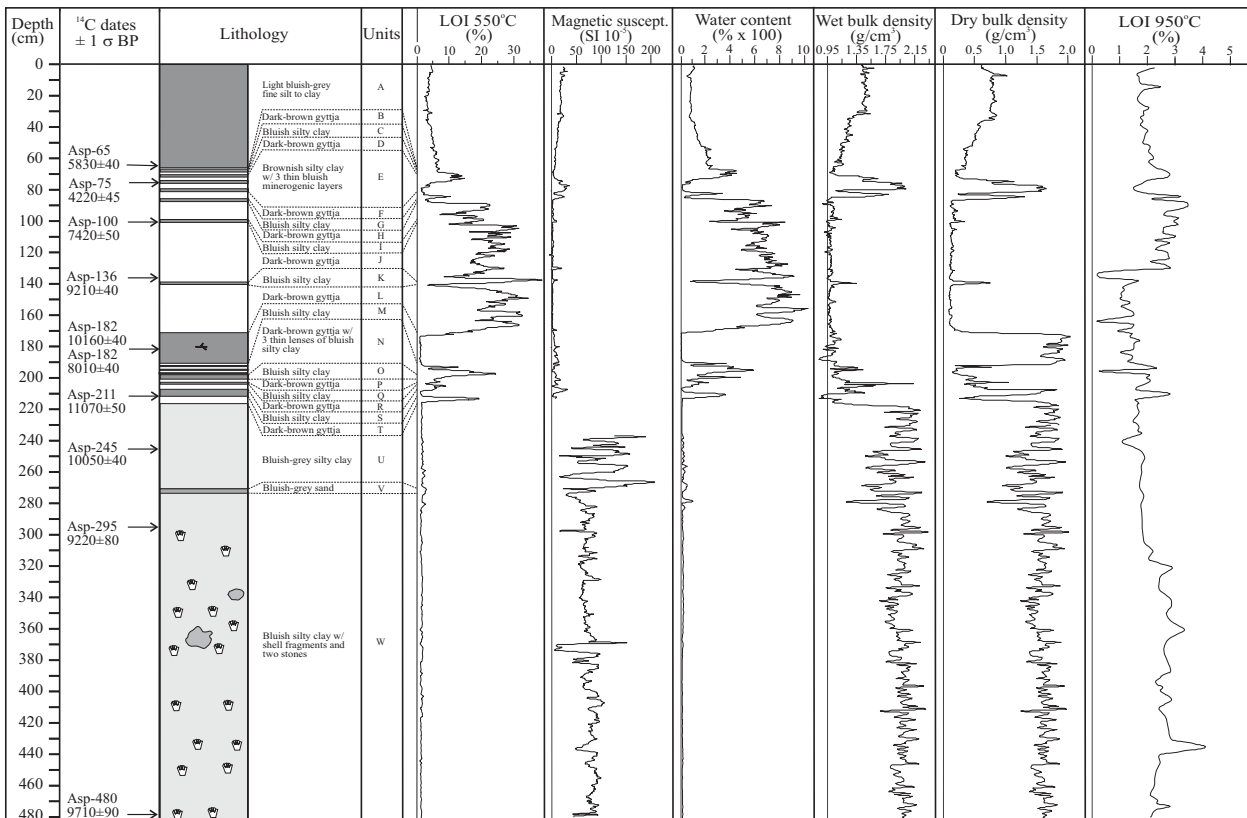
the fan. Hence, the meltwater stream from Strupskardet is suggested to have followed this course during the Holocene. Only a small portion of the glacier meltwater entering this fan, however, ends up in Aspvatnet (Figs. 4 and 10).

### Lithostratigraphy and radiocarbon dating

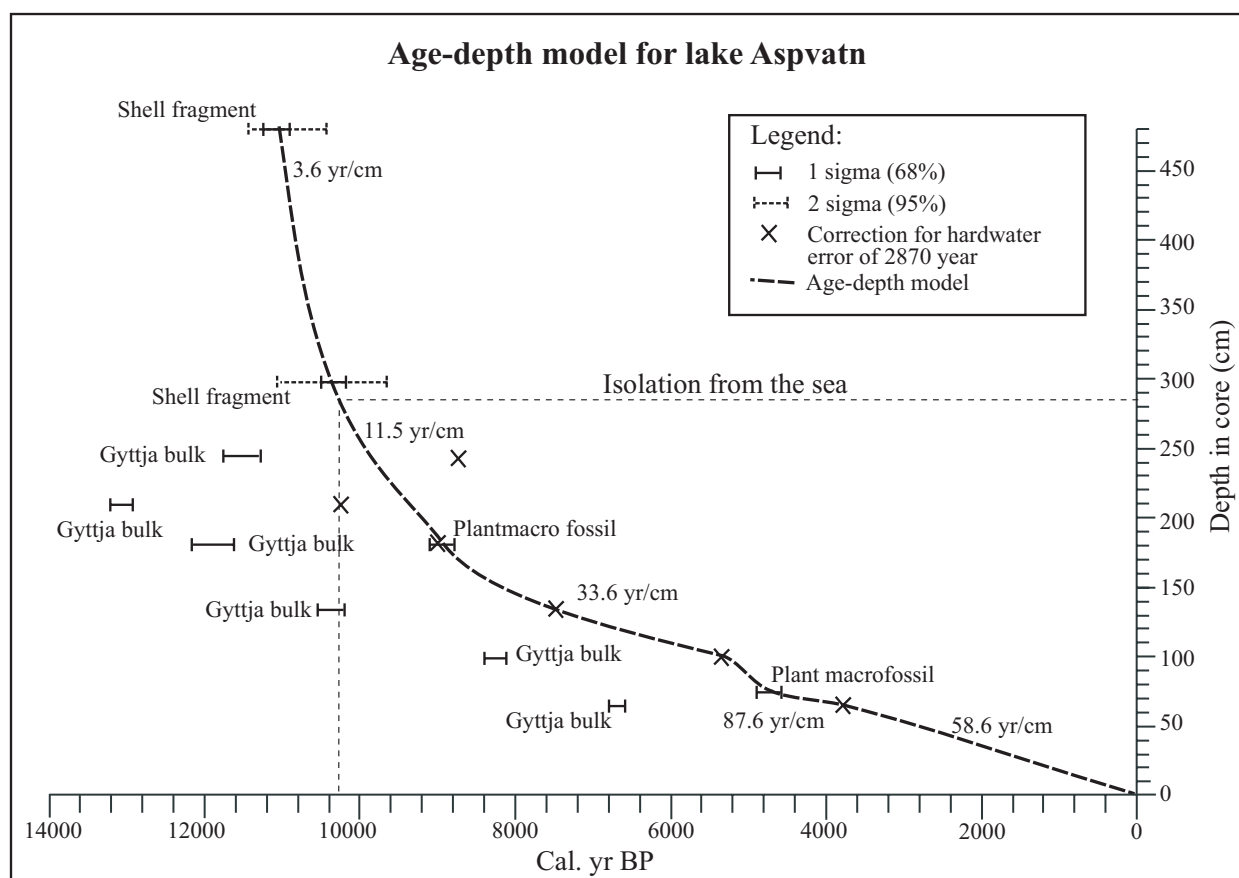
Based on visual description and a range of different sediment analyses, the lithostratigraphy in core 1 has been subdivided into 23 (A-W) units (Fig. 12). To produce an age-depth model, ten accelerator mass spectrometry (AMS) radiocarbon dates of shell fragments, terrestrial plant macrofossils and bulk samples from core 1 (Table 1) were obtained and calibrated to calendar years using INTCAL98 (Stuiver et al., 1998).

The age-depth model is based on linear interpolation between the dated levels and by assuming that the uppermost sample is modern (Fig. 13). Because of a nearly constant hardwater effect, some of the corrected bulk samples were included in the age-depth model. The two cores from Aspvatnet is correlated and highly consistent (Fig. 14).

The age-depth model and the environmental interpretation of the two cores from Aspvatnet are summarized in Figure 14.



**Figure 12** AMS-radiocarbon dated lithostratigraphy from core 1 in Aspvatnet with curves describing variations in loss-on-ignition (LOI) at 550 °C, magnetic susceptibility (MS), water content (in % of dry weight), wet bulk density (g/cm<sup>3</sup>), dry bulk density (g/cm<sup>3</sup>) and LOI at 950 °C. Because of problems with the surface sensor, there is a break in the magnetic susceptibility measurements at 220 cm.



**Figure 13** Age-depth model for core 1 in Aspvatnet based on the obtained AMS radiocarbon dates shown in Table 1. The two lowermost radiocarbon dates are obtained from marine shells and are corrected for a marine reservoir age of 400  $^{14}\text{C}$  yrs.

## Discussion

### The age of the marginal moraines in Strupskardet

The ice-cored moraines M1-M3 deposited by polythermal glaciers and the more-or-less contemporaneous talus-derived rock glaciers indicate a continuous permafrost regime in Strupskardet during the formation. A sub-marine formation of M1-M3 is regarded unlikely, and the corresponding sea level must therefore have been lower than 20 m for the ice-cored moraine complex to develop. Such a relatively low global sea level is inferred to have existed during the time span of maximum land-based ice volume from 22,000 to 19,000 cal. yr BP (Yokoyama et al., 2000). The  $L_4$  sea level correlated to the Risvik substage (c. 20,000 –

18,000 cal. yr BP) in Porsangerfjorden in Finnmark (Sollid et al., 1973; Andersen, 1979) have abraded M1-M3. Hence, the age of these moraines must be somewhat older or from this time span (Figs. 4, 7 and 8).

The M4-M7 moraines are suggested to be push-and melt-out moraines formed by advancing glaciers. A relict meltwater channel linked to M4 terminates in the  $L_4$  sea level at 87 m a.s.l. (Fig. 4), whereas the M5-M7 moraines are linked to a relict glaciofluvial fan abraded by the  $L_1$  sea level at 67 m a.s.l. correlated to the Skarpnes event (c. 14,500-14,000 cal. yr BP) (Figs. 4, 7 and 8).

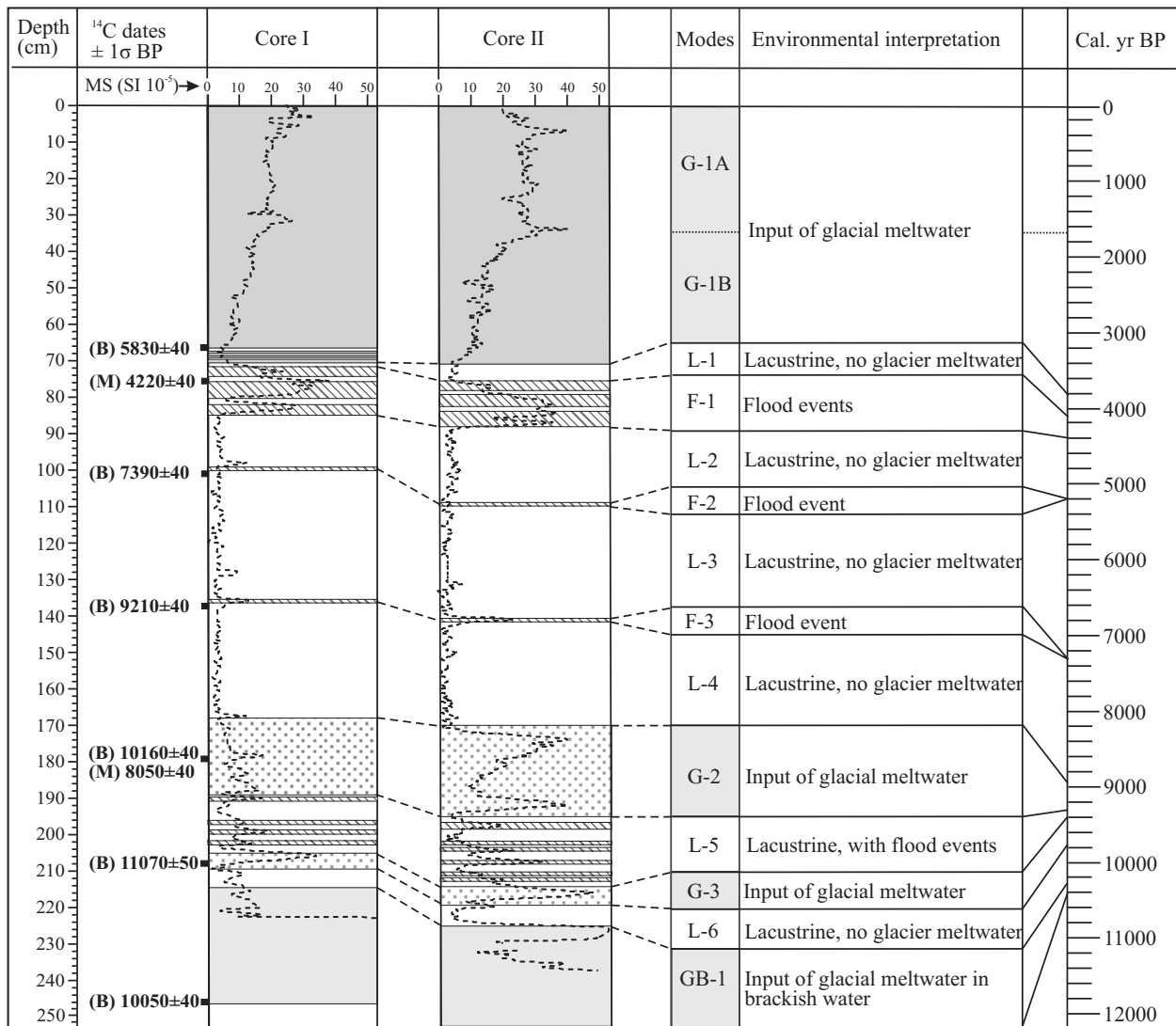
The suggested relict ice-cored moraines M8 and M9 deposited by polythermal glaciers and the occurrence of contemporaneous frost polygons indicate continuous permafrost conditions when these features were formed. A former meltwater

**Table 1** Radiocarbon dates from core 1 in Aspvatnet. When more than two possible intercept ages, the mean value was used.

Site	Lab. no.	Depth (cm)	Type of material	Conventional $^{14}\text{C}$ age BP	$\delta^{13}\text{C}$ (‰)	$\pm 1 \sigma$ Cal. yr BP	$\pm 2 \sigma$ Cal. yr BP	Intercept Cal. yr BP
Aspvatn	Beta - 154058	66-65	Gyttja	5830 $\pm$ 40	-25.0	6670-6630	6730-6530	6650
Aspvatn	T-8987	75.5	Macrofossil	4220 $\pm$ 45	?	4845-4650	4860-4585	4700
Aspvatn	Beta - 154059	100-99	Gyttja	7420 $\pm$ 50	-23.3	8330-8180	8350-8160	8200
Aspvatn	Beta - 154060	136-135	Gyttja	9210 $\pm$ 40	-25.2	10,420-10,260	10,500-10,240	10,310
Aspvatn	Beta - 154061	182-180	Gyttja	10,160 $\pm$ 40	-24.7	12,110-11,680	12,290-11,580	11,860
Aspvatn	Beta - 154827	182-180	Macrofossil (wood)	8010 $\pm$ 40	-27.5	9000-8790	9020-8740	8990
Aspvatn	Beta - 154062	211-210	Gyttja	11,070 $\pm$ 50	-25.0	13,160 – 12,910	13,180-12,890	13,090
Aspvatn	Beta - 154063	245	Gyttja (macro)	10,050 $\pm$ 40	-24.8	11,910- 11,340	11,940-11,300	11,560
Aspvatn	Beta - 154064	295	Shell	9220 $\pm$ 80*	2.2	10,480-10,245	10,575-10,220	10,400
Aspvatn	Beta - 154065	480-478	Shell	9710 $\pm$ 90*	-0.5	11,200-10,790	11,255-10,740	10,995

\*Corrected for marine  $^{14}\text{C}$  reservoir age of 400 years (Bondevik et al., 1999)





**Figure 14** Composite log showing the lithology and environmental interpretation based on cores 1 and 2 in Aspvatnet. Four periods with input of glacier-meltwater induced sediments have been inferred. Three river floods have been identified in the lake record. There is good correlation between the two cores except for some minor discrepancies between sedimentation rates and lithology.

channel distal to M8 and M9 along the northern valley slope of Strupskardet links these marginal moraines to the  $L_0$  sea level from the Younger Dryas (c. 12,800-11,500 cal. yr BP) (Figs. 4, 7 and 8).

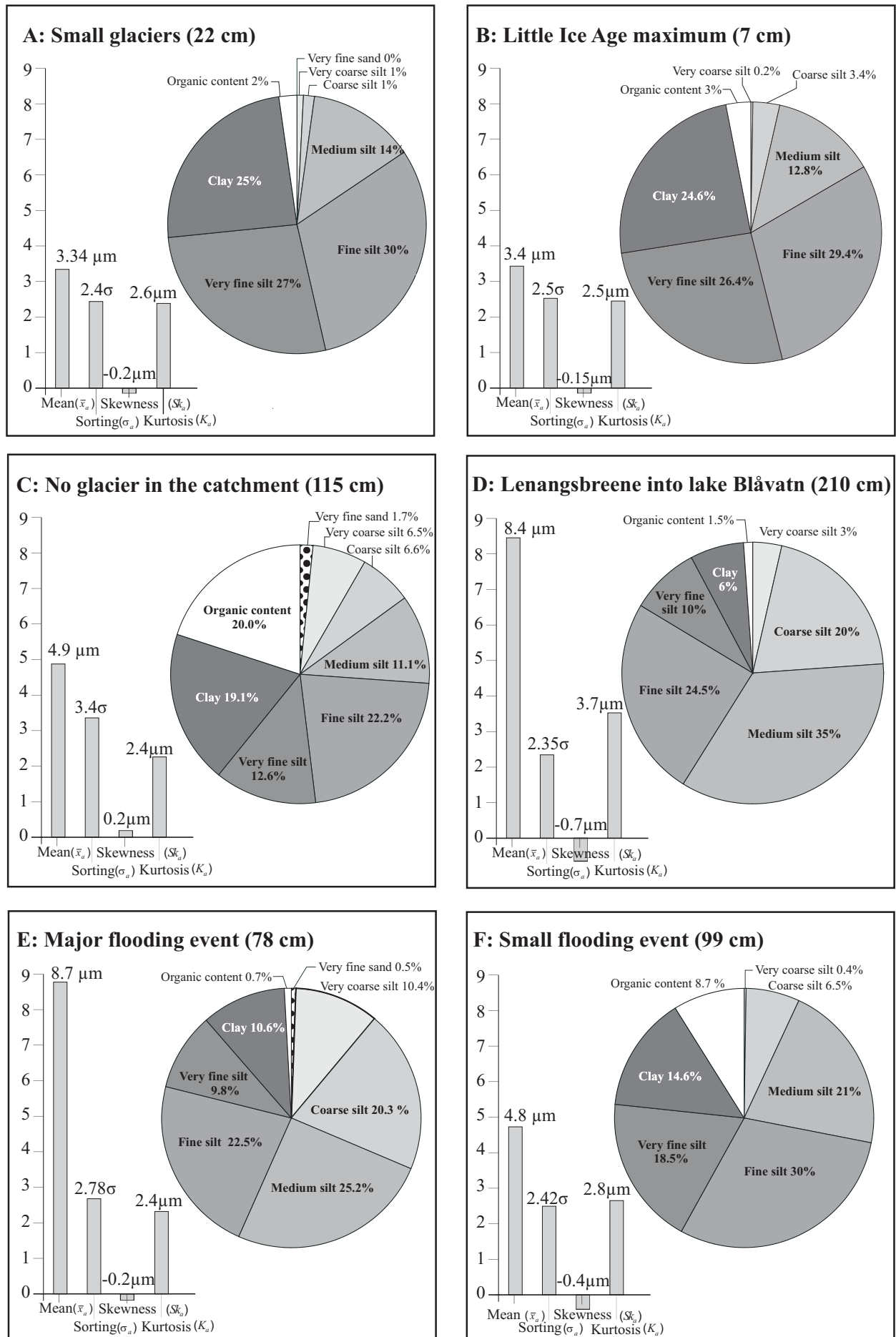
The push- and melt-out moraines M10, M11 and M12 are deposited by advancing glaciers, and are indirectly dated by AMS radiocarbon dates from lacustrine sediments in Aspvatnet at 10,400-10,300, 9800-9400 and 9300-8900 cal. yr BP, respectively (Fig. 14) (see below for further discussion). The push moraine M13 is dated by use of lichenometry to AD 1890-1928 (Fig. 9), and is suggested to represent the 'Little Ice Age' glacier maximum in Lyngen (e.g. Ballantyne, 1990).

#### Sediment analyses from Aspvatnet

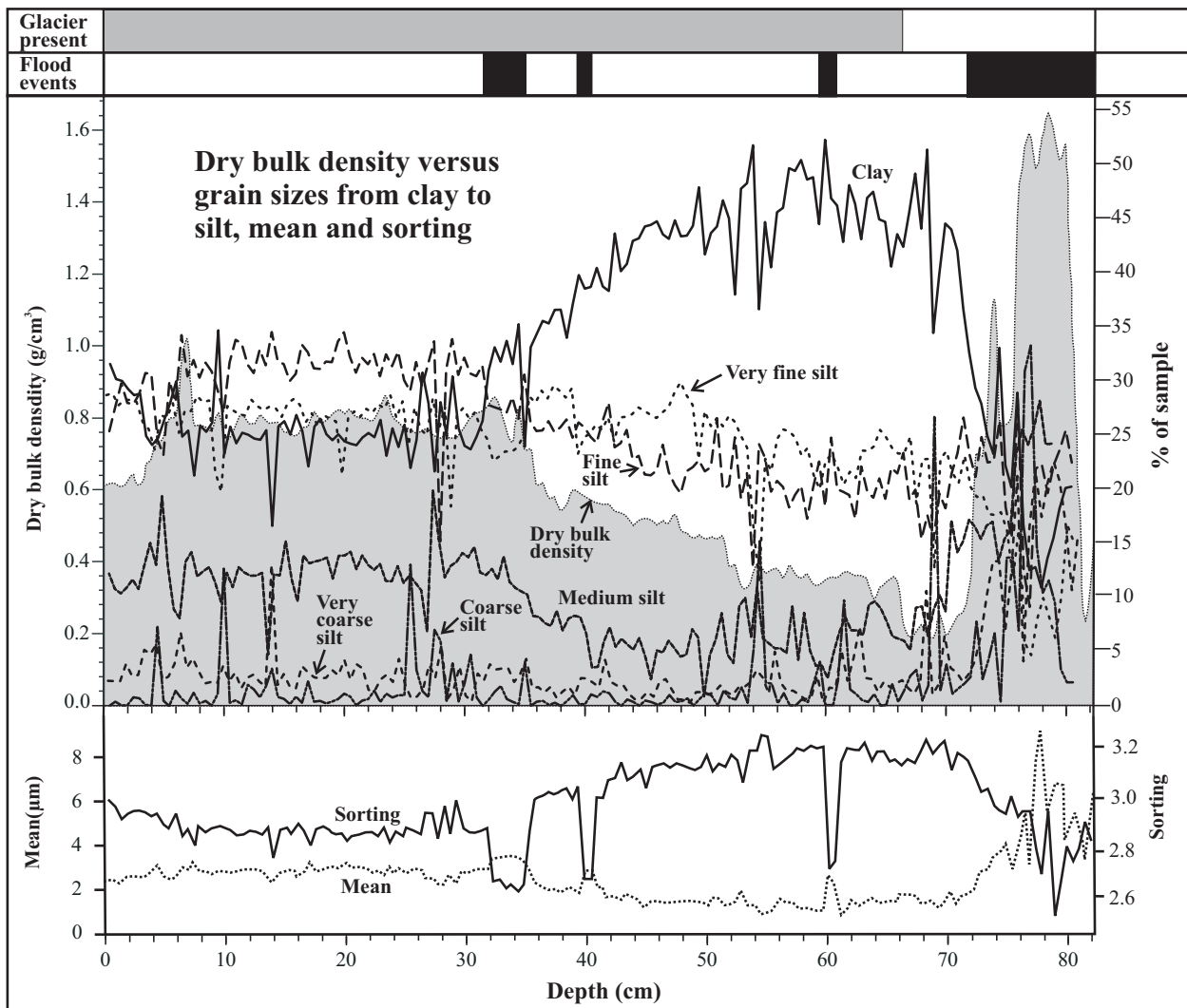
Adjusted for a marine reservoir age of 400 years, the sedimentation in Aspvatnet started around 11,100 cal. yr BP according to the basal date on shell fragments (Table 1). Hence, Aspvatnet can not be regarded as a lacustrine site before the isolation from the sea around 10,300 cal. yr BP (Fig. 8) (Corner and

Haugane, 1993). The marine phase was characterised by high sedimentation rates (mean 3.6 yr/cm), and it is assumed that there was input of glacial meltwater to Lenangen (unit W in Fig. 12). Unit V marks the first isolation from the sea with a layer of fine sand, and the start of a period with brackish water and continuous input of glacier-induced sediments (unit U in Fig. 12, mode GB-1 in Fig. 14). Unit U culminated by a final isolation from the sea and transition to lacustrine conditions with no glaciers in the catchment in unit T (Figure 12, mode L-6 in Figure 14).

During the early-Holocene lacustrine phase in Aspvatnet, two events with glaciers in the catchment are recorded in units S and M in Figure 12, corresponding to modes G-3 and G-2 in Figure 14. In between these modes, Aspvatnet was dominated by lacustrine sediments interrupted by river-floods in mode L-5. The marginal moraines M10, M11 and M12 in Strupskardet (Figs. 4 and 6) are correlated to modes GB-1, G-3 and G-2 in Figure 14, and are AMS radiocarbon dated on lacustrine sediments in



**Figure 15** Six selected samples from core 1 in Aspvatn showing grain-size distribution, mean, sorting, skewness and kurtosis. Sediment sample A is from mode G-1A (small glacier), B is from mode G-1B (large glacier), C is from mode L-3 dominated by gyttja, D is from mode G-3 dominated by glacially derived sediments during a time span when Lenangsbreen calved into Blåvatnet, E is from mode F-1 suggested to be a major river flood and G is from mode F-2 suggested to be a minor river flood.



**Figure 16** In the upper panel, the dry bulk density (DBD) from 0-82 cm depth in core 1 from Aspvatnet is shown as a grey-shaded area, whereas the grain-size distribution of clay and various silt fractions are presented in per cent of each sample. In the lower panel 'sorting' and 'mean' based on the grain-size analysis are shown. Four prominent 'mean-sorting anomalies' suggested to represent river floods are found at 82-74, 60, 40 and 34 cm depth in the core. Note that the relative content of clay decreases in the uppermost 40 cm of the core. The high dry bulk density values from 82-74 cm are from mode F1 in Figure 14, and are suggested to represent a major river flood.

Aspvatnet to 10,400-10,300, 9800-9400 and 9300-8900 cal. yr BP, respectively. From about 8800 to 3800 cal. yr BP there was no input of glacier-meltwater induced sediments to Aspvatnet, and the regular lacustrine sedimentation in the lake was only interrupted by large flooding events during modes F-3, F-2 and F-1 in Figure 14 (see below for further discussion).

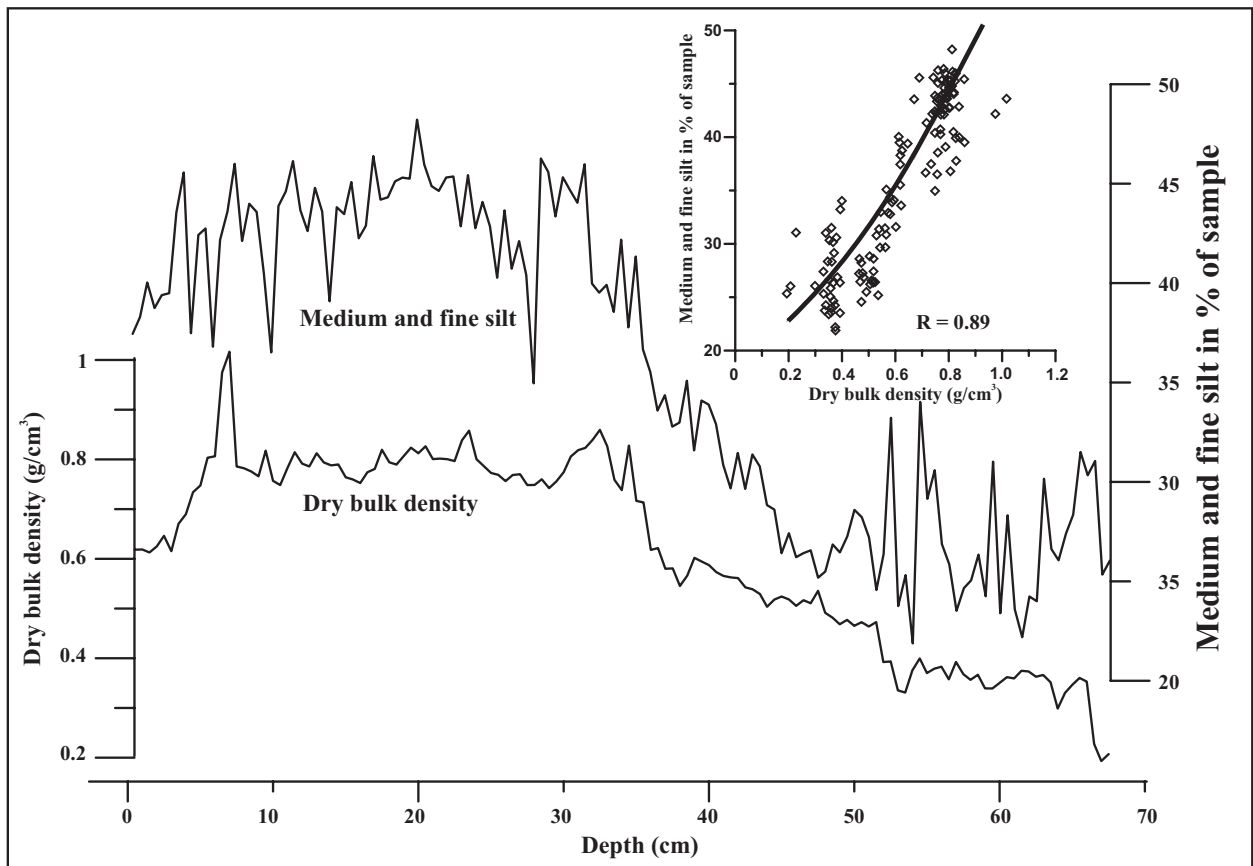
#### Dry bulk density and grain-size analysis

Glacial erosion supplies bedrock particles to glacier-meltwater streams, and analysis of particle size may be a diagnostic parameter for glacier size (Matthews et al., 2000; Lie et al., in press). Wet-based, temperate glaciers produce abundant clay-silt size fractions that are transported downstream to produce characteristic signatures in glacio-lacustrine sediments (e.g. Østrem, 1975). The use of grain-size variations have, however, not been widely used in this context. One problem with the parameters is that it is difficult to flux correlate the signal, as the

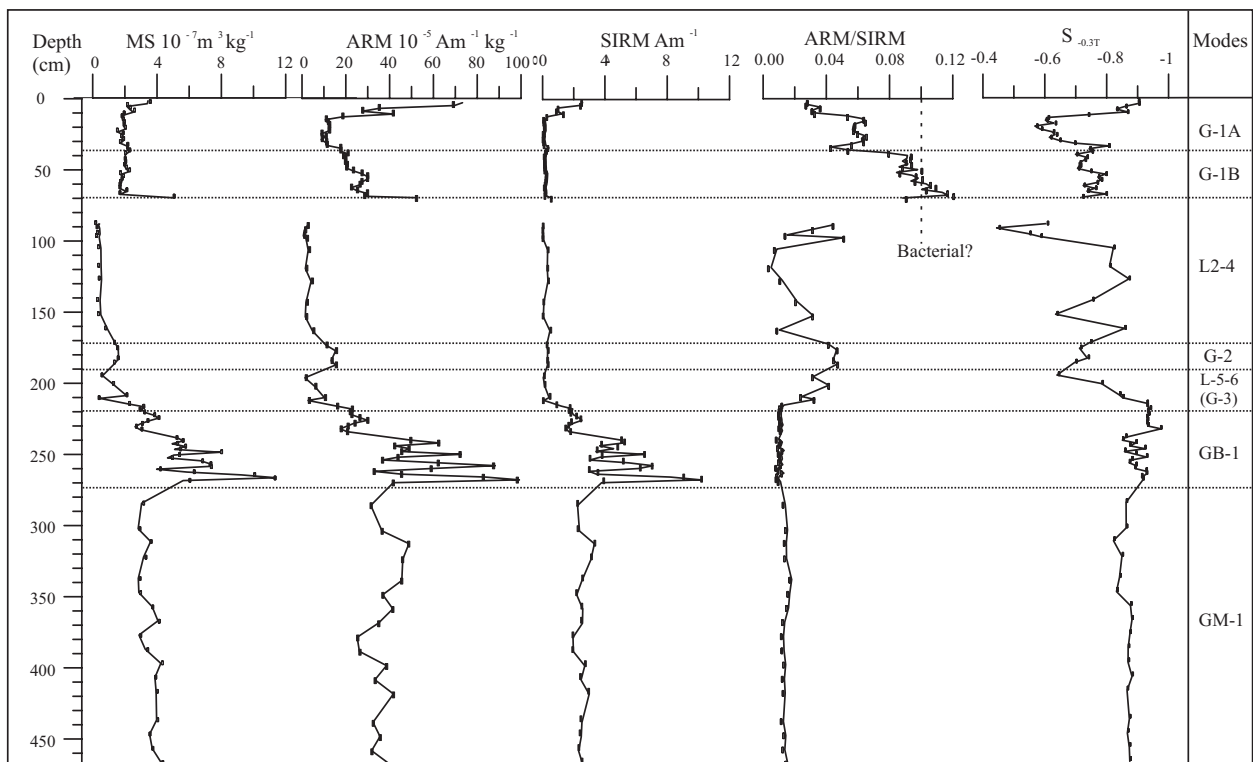
relatively long time intervals between the radiocarbon dates are a limiting factor. The distribution of grain sizes in proglacial lake sediments is mainly an indicator of the energy of the water masses flowing through the lake. As the sedimentation is closely related to Hjulström's diagram (Sundborg, 1956) of transportation in fluvial systems, it is apparent that high energy gives less of the finest grain sizes, and *vice versa*.

The grain-size analyses from core 1 in Aspvatnet have been used to establish time intervals with stable sedimentation and periods dominated by sedimentation from suspension. Based on the analysis, there is also established a relationship between grain size and dry bulk density related to glacier size. From this, six different Holocene lacustrine sedimentation environments have been recognized in Aspvatnet and shown in Figure 15: (1) meltwater input from small glaciers, (2) meltwater input from large glaciers ('Little Ice Age maximum'), (3) organic-dominated sedimentation with no glacier

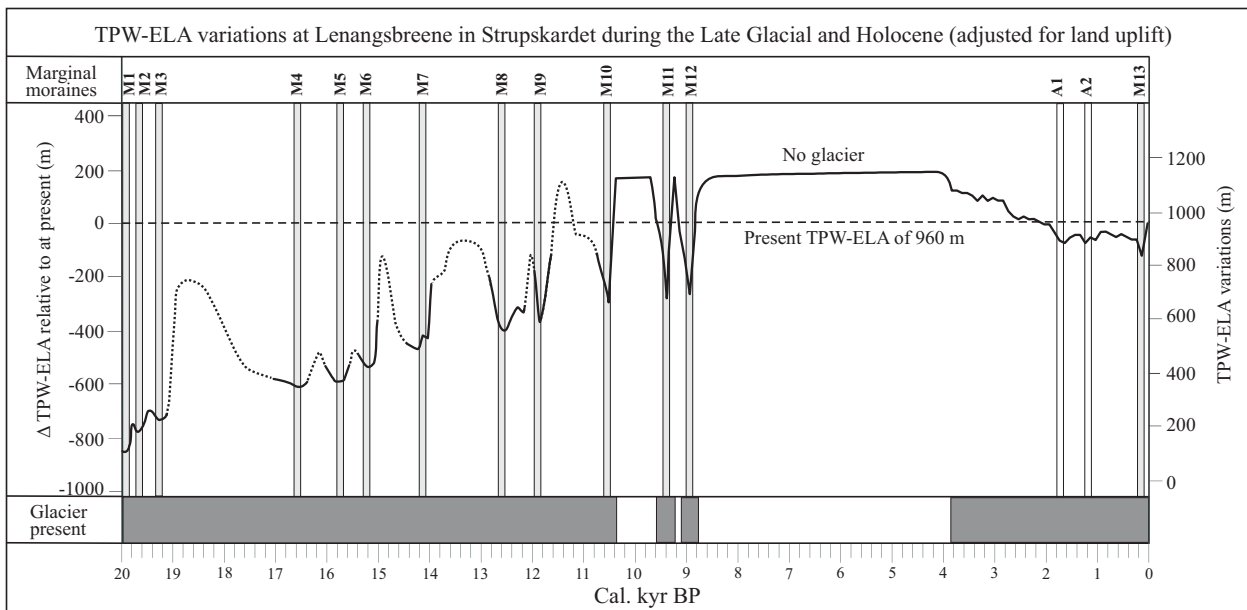




**Figure 17** The main figure shows variations in a combination of fine and medium silt compared with fluctuations in dry bulk density (DBD) in the upper part of core 1 from Aspvatnet, whereas the incised figure shows the correlation ( $r=0.89$ ) between dry bulk density and a combination of fine- and medium silt. For ‘open-ended’ lakes such as Aspvatnet, it is suggested that the amount of finer grain sizes transported out of the lake is positively correlated with variations in the amount of water that enters the lake. The inferred ‘Little Ice Age’ (LIA) maximum and the glacier advances A1 and A2 without preserved marginal moraines are marked on the diagram.



**Figure 18** Diagram shows variations in environmental magnetic parameters in core 1 from Aspvatnet. Note the section with possible magnetotactic bacteria in the lower part of mode G-1B. Sample intervals vary with depth in the core. See text for further discussion.



**Figure 19** Mean TPW-ELA of Lenangsbreene in Strupskardet from 20,000 cal. yr BP to at present. The TPW-ELAs are reconstructed using an AAR of 0.6 on the moraines M8 and M13 and a technique termed the 'Little Ice Age ratio' (LR) (Dahl *et al.*, 2002) adapted to Strupskardet for the remaining marginal moraines (See Table 2 and text for further discussion). The reconstructed TPW-ELAs are adjusted for land uplift according to the sea-level history presented in Figure 8.

in the catchment, (4) very large glaciers (Lenangsbreene calving in Blåvatnet), (5) large high-energy river floods, and (6) small low-energy river floods.

During the early-Holocene glacier events that formed the moraines M10-M12 (Fig. 4), sedimentation in Aspvatnet was dominated by group 4 (mode GB-1, G-3 and G-2) and group 3 (mode L-6, L-5 and L-4) with abrupt changes between clastic sedimentation and gyttja (Figs. 14 and 15). However, two (three) episodes during the early- and mid-Holocene are recognised as small low-energy river floods belonging to group 6 (mode F-3 and F-2), and a major flooding event, representing group 5 (mode F-1), occurred from 85-73 cm in core 1 (Figs. 12 and 14). Group 1 is recognised as small glaciers

(mode G-1B), whereas group 2 represents large glaciers ('Little Ice Age' maximum) (mode G-1A) in the upper part of core 1 linked to moraine M-13 (Figs. 4, 12, 14 and 15).

Variations in grain size and dry bulk density are shown in Figure 16. Dated to c. 4000 cal. yr BP, the high dry bulk density values between 85-73 cm represent the major flooding event described in group 5 (mode F-1) in Figures 14 and 15. After c. 3800 cal. yr BP (65 cm in Fig. 16), however, Aspvatnet has been characterised by stable sedimentation of glacially derived sediments from suspension. A lowering in the relative proportion of the finest grain sizes and the simultaneous rise in coarser grain sizes above 35 cm are suggested to represent increased mean fluvial energy as Lenangsbreene became larger.

**Table 2** Marginal moraines M1-M13 with dating method and the corresponding observed and relative TPW-ELA changes adjusted for land uplift.

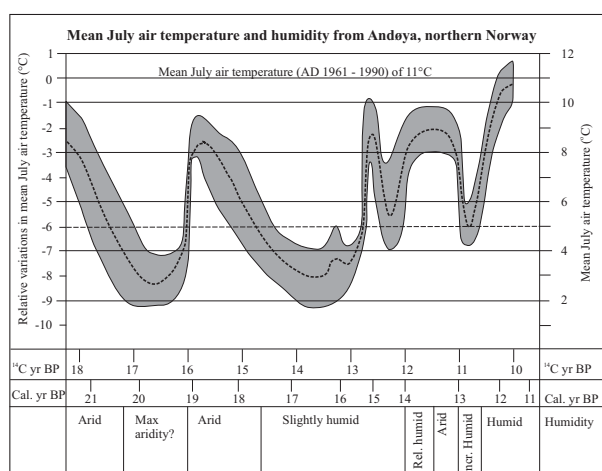
Marginal moraine	Cal. yr BP (AD 1950 = 0)	Sea-level (m a.s.l.)	TPW-ELA (m)	TPW-ELA (m) (adjusted)	$\Delta$ TPW-ELA (m) (adjusted)	Dating method
M1	20,000	+ 20	140	120	- 840	Sea level
M2	19,800	+ 20	210	190	- 770	Sea level
M3	19,200	+ 35	280	245	- 715	Sea level
M4	16,600	+ 80	435	355	- 605	Sea level
M5	15,700	+ 75	445	370	- 590	Sea level
M6	15,200	+ 70	490	420	- 540	Sea level
M7	14,200	+ 65	560	495	- 465	Sea level
M8	12,500	+ 60	630	570	- 390	Sea level
M9	11,700	+ 57	635	580	- 380	Sea level
M10	10,500	+ 38	700	660	- 300	<sup>14</sup> C dating
M11	9400	+ 30	705	675	- 285	<sup>14</sup> C dating
M12	9000	+ 29	720	690	- 280	<sup>14</sup> C dating
M13	AD 1910	0	840	840	- 120	Historical/ Lichenometry
Present TPW-ELA	AD 2003	0	960	0	0	Observed

This is accompanied by higher dry bulk density interpreted as higher sediment flux, which is also supported by changes in the magnetic parameters (Fig. 18).

A combination of the two parameters 'sorting' and 'mean' ( $\mu\text{m}$ ) gives additional information regarding the sedimentation environment in Aspvatnet (Fig. 16). Sorting reflects the steepness of the cumulative grain size distribution curve in the sample, and the parameter 'mean' gives the corresponding mean grain size. Periods with stable sedimentation were characterized by a gradual rise or fall in both parameters as all changes appear gradually. Periods when the parameters were in strong and abrupt anti-phase is suggested to represent rapid catastrophic events, such as river floods indicated by reduced 'sorting' accompanied by increased particle size (Arnaud *et al.*, 2002). When sorting is gradually lowered and the mean is gradually rising, this may indicate a more long-lasting change in the sedimentation environment such as a general increase in the summer runoff from a larger glacier. By combining 'sorting' and 'mean' in Figure 16, river floods are suggested to have taken place at 85-73, 60, 40 and 35-32 cm in the investigated core 1. As the deposited layers identified as flooding events have the properties of glacier-meltwater derived sediments based on visual description, LOI and DBD, the sediments may be regarded as paraglacial material (Church and Ryder, 1972). The same argument may be used for the deposition of the modes F1-F3 in Figure 14.

After removing the river-flood signals, a continuous record of glacier variations is suggested to be reflected by a combination of medium to fine silt, and by the flux parameter dry bulk density (Fig. 17). The correlation between dry bulk density and medium to fine silt is 0.89. This high correlation is suggested to be a reflection of Aspvatnet as an open-ended lake (e.g. paper 3) where the inlet and outlet are located so close to each other that the amount of fine material in suspension transported out again reflects the residence time of water in the lake.

Suggested to be representative for glacier variations for the last 3800 cal. yr BP, the dry bulk density curve in Figure 17 has been calibrated to reflect a continuous ELA curve for this time span. The calibration is done based on samples from Aspvatnet reflecting the present situation, the 'Little Ice Age' glacier maximum with formation of M13, and periods when the glacier was melted away. Marginal moraines from inferred glacier advances at 32 and 23 cm in Figure 17 are named A1 and A2. If these dry bulk density events are glacier advances, corresponding marginal moraines were later erased by the 'Little Ice Age' maximum glacier, as probably reflected at 7 cm in Figure 17.



**Figure 20** Reconstructed mean July air temperature and humidity at Andøya, northern Norway. The dotted line is used as temperature estimate during the Lateglacial for quantification of former winter precipitation in Lyngen. Modified from Vorren *et al.* (1988).

### Magnetic modes

Variations in magnitude and pattern of sediment parameters, divide the core into seven modes/zones (Fig. 18) that are in general agreement with the described lithostratigraphy in Figure 12 and the environmental interpretation in Figure 14:

- Mode GM-1 (469-270cm, N=18) represents a marine phase (Fig. 12) with input of glacier-derived meltwater and shows relatively uniform values associated with small-amplitude variations of concentration dependent parameters. S-0.3T ratios close to -1.0 indicates dominance of ferromagnetic minerals. The magnetic grain-size indicator (ARM/SIRM) reflects larger than single domain (SD) grains.

This zone may consist of sediments derived from a single source. However, a similar constant signature may also be found in sediments derived from several sources that have been mixed by depositional processes like tidal currents, bioturbation etc.

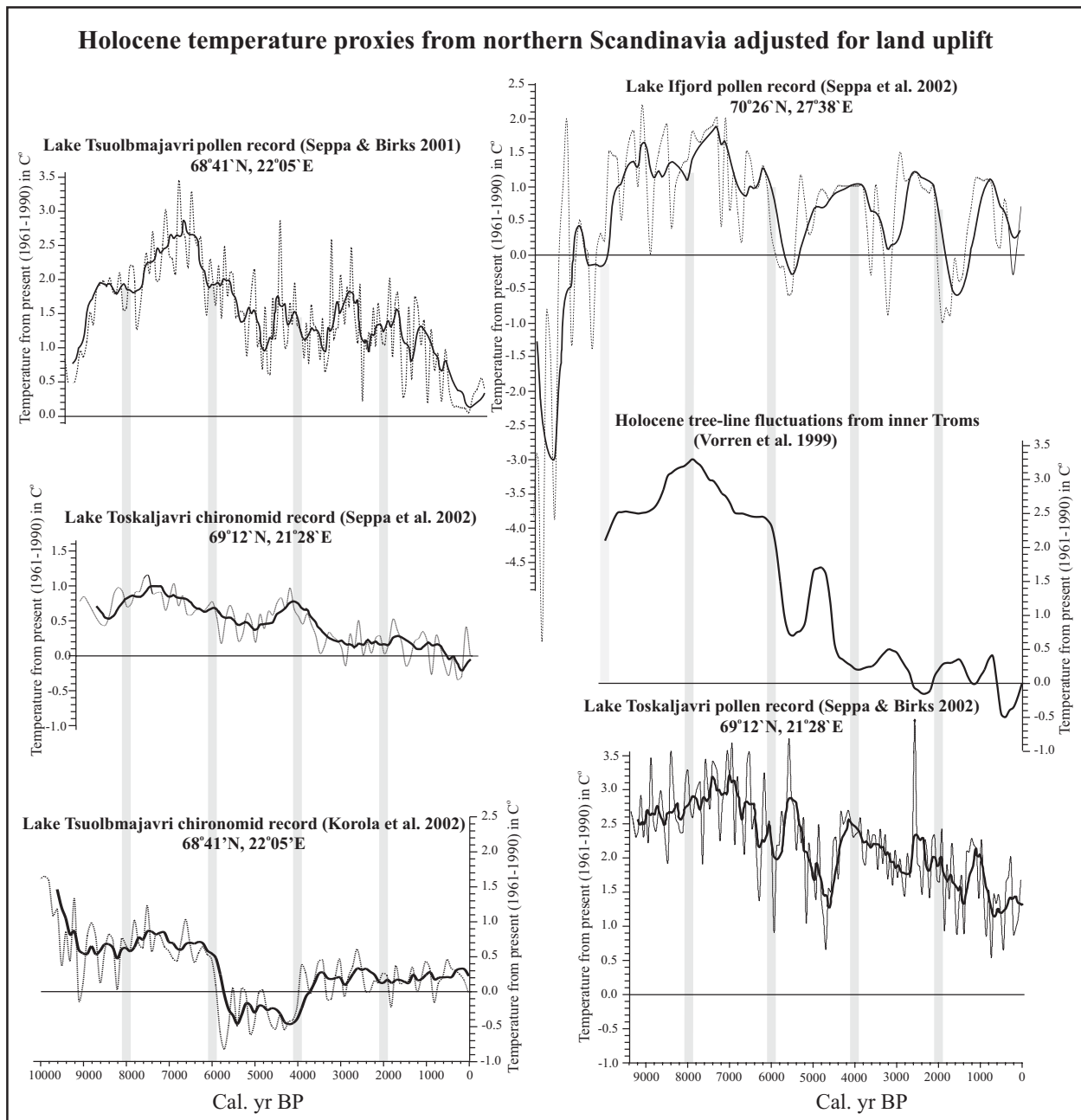
- GB-1 (268-216 cm, N=25) represents a brackish phase with input of glacier-derived meltwater and has S-0.3T values close to -1.0 (magnetite) and low ARM/SIRM ratios indicative of larger than SD grains. Concentrations of magnetic minerals show a rapid decrease accompanied by high-frequency variations.

This indicates a gradual decrease in the amount of magnetic minerals from the source(s), and the absence of effective mixing processes.

- L-5-6 (G-3) (212-173 cm, N=8) represents a lacustrine phase with no glaciers in the catchment, and has fairly constant amounts of magnetic minerals. A small systematic increase in ARM/SIRM indicates a 'fining upward' of magnetite grains. A single anomalously low S-0.3T-value may imply a temporary source.

- L-2, L-3 and L-4 (163-70 cm, N=11) represent lacustrine phases with no glaciers in the catchment, and have a uniform concentration of magnetic minerals as judged from MS, ARM and SIRM. Large





**Figure 21** Five selected continuous temperature reconstructions using biological proxies from lakes and one using pine-tree limit fluctuations in northern Scandinavia are adjusted for land uplift. The reconstructions, taken into account vegetation changes from the lakes Tsuolbmajavri (Seppä and Birks, 2001), Ifjord (Seppä *et al.*, 2002a) and Toskaljavri (Seppä and Birks, 2002), are all based on pollen transfer functions, whereas the reconstructions based on non-biting midges from the lakes Tsuolbmajavri (Korhola *et al.*, 2002) and Toskaljavri (Seppä *et al.*, 2002b) are based on transfer functions using chironomids. The temperature reconstruction using variations in the pine-tree limit is from inner Troms (Alm, 1993; Vorren *et al.*, 1996). As there are major discrepancies from one site to the next and from one proxy to another, a mean of all temperature reconstructions are used for quantifying winter precipitation during the last 4000 cal. yr BP in Lyngen (see Figure 22).

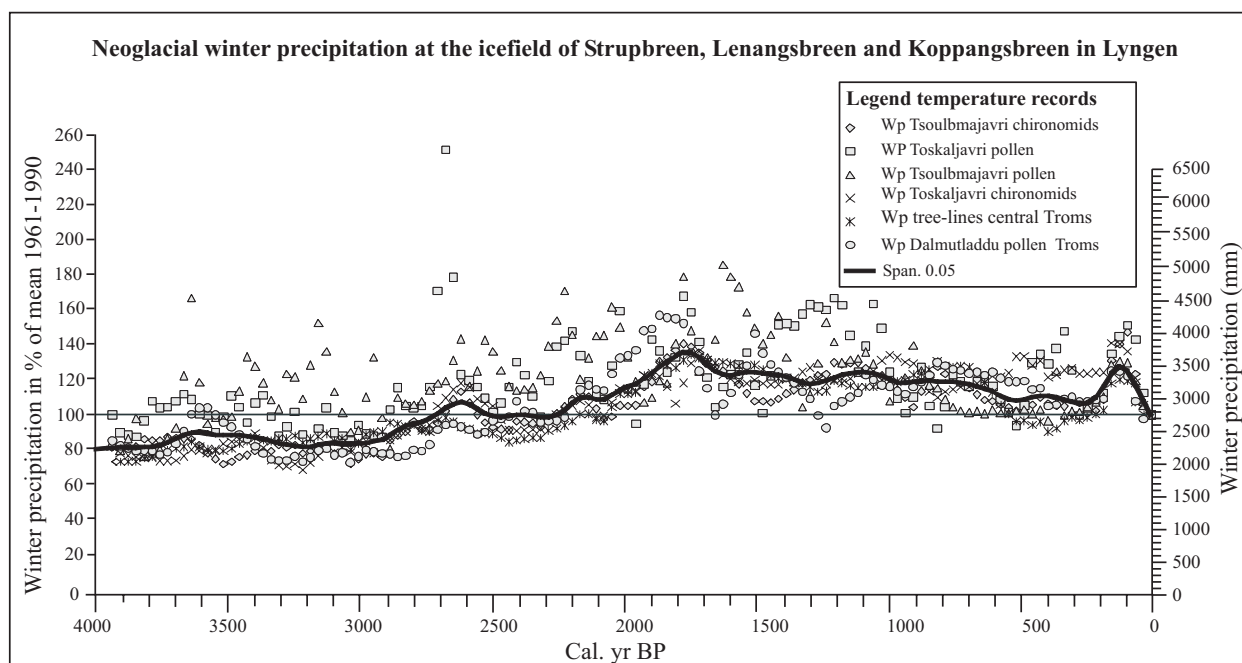
variations in magnetite grain size and S-0.3T suggest significant changes in composition implying no effective mixing processes and several sources.

- G-1B (68-34 cm, N=18) represents a phase with small glaciers in the catchment, and shows a constant concentration of large magnetite (MS and SIRM). ARM and ARM/SIRM show an almost linear decrease in the amount of SD grains, and ARM/SIRM ratios above 0.1 may reflect the dominance of bacterial magnetite in an anaerobic environment (Moskowitz *et al.*, 1993).

- G-1A (32-12 cm, N=11) represents a phase with

small glaciers in the catchment, and has a constant mixture of SD and larger grains. ARM/SIRM is significantly higher than below mode L-2, L-3 and L-4. Since an ARM/SIRM ratio less than 0.1 does not exclude the presence of bacterial magnetite, magnetosomes may be present in this zone, but are diluted with larger, detrital magnetite grains. With S-ratios significantly below -1.0, the occurrence of bacterial magnetite is uncertain.

- G-1A (10-2 cm, N=5) represents a phase with large glaciers (the 'Little Ice Age' maximum) in the catchment, and consists of increasing amounts of



**Figure 22** Mean winter precipitation at Lenangsbreen from 3800 cal. yr BP to at present. The thick line is a lowess filter with a span of 0.05 which gives the average winter precipitation estimate based on calculations using the different temperature reconstructions in Figure 21. In addition, an unpublished pollen-based temperature reconstruction from Dalmutladdu in inner Troms (Bjune *et al.*, in press) is used.

magnetic material reflected by increasing values of MS, ARM and SIRM. ARM increases about 5 times across this zone and SIRM experiences an almost 10 times increase associated with ARM/SIRM ratios below 0.1. The large increase in ARM indicates increasing concentration of small magnetite grains. As the ARM/SIRM ratio is well below 0.1, however, it is tentatively concluded that bacterial magnetite may be present but that the magnetic signature is masked by coarser magnetite grains.

### Lateglacial and Holocene variations in magnitude and TPW-ELA

The various glacier events in Strupskardet have been dated based on the distribution of permafrost features, the connection between the various marginal moraines and the regional sea-level history, the AMS radiocarbon-dated lacustrine sequence in Aspvatnet and lichenometry.

The M8 glacier is well-defined by use of the marginal moraines, and the M13 glacier is well-defined based on lichenometry and local topography in front of Lenangsbreen (Figs. 4 and 6). The reconstructed glacier profiles of M8 and M13 are relatively steep and they are both suited for the use of AAR to calculate former TPW-ELAs. By use of an AAR of 0.6, the former TPW-ELAs of the M8 and M13 glaciers were -330 and -120 m, respectively. The marginal moraines beyond M8, including M1-M3, are located on relatively flat ground and thus less suited for use of AAR (e.g. Nesje and Dahl, 1991). A new technique termed the 'Little Ice Age ratio' (LR) (Dahl *et al.*, 2002) adapted to Strupskardet has

therefore been used to estimate former TPW-ELAs on the remaining marginal moraines. The distance between the termini of M13 and M8 is 1500 m and corresponds to a drop in the TPW-ELA of 210 m. By using this ratio (ELA lowering (m)/distance (m)) not adjusted for land uplift, corresponding TPW-ELAs were estimated on the remaining marginal moraines, and shown in Table 2 and Figure 19 adjusted for land uplift.

Accurate estimates for the timing and amplitude of the TPW-ELA when the glacier is retreating involves an uncertainty, and both the temperature reconstruction from Andøya (Vorren *et al.*, 1988) and the event stratigraphy from the Andfjord-Vågsfjord area in northern Norway (Vorren and Plassen, 2002) have been taken into account. However, most of the glacier advances are suggested to be shorter halts in a general retreat of the glaciers in Strupskardet.

### TPW-ELAs during the Lateglacial before c. 18,000 cal. yr BP

The relict ice-cored moraine complex (M1-M3) in Strupskardet is assumed to have developed while the sea level was lower than 20 m a.s.l during the time span of maximum global land-based ice volume from 22,000 to 19,000 cal. yr BP (Yokoyama *et al.*, 2000), and before this complex was abraded by the L<sub>4</sub> sea level during the Risvik substage no later than c. 20,000 - 18,000 cal. yr BP. This is also comparable with the glaciation diagram from Andfjord-Vågsfjord, where the glacial events Bjerka and Egga II are placed with Bennin this time interval (Vorren and Plassen, 2002). The rapid global decrease in land-based ice volume after 19,000 cal. yr BP (Yokoyama *et al.*, 2000) with release of water to the sea, fits well

with the assumed age of the  $L_4$  sea level in Lyngen with an estimated altitude of c. 87 m in Strupskardet (Figs. 7 and 8).

Adjusted for a land uplift of 20 m for M1 and M2, the corresponding lowering of the TPW-ELAs were 840 and 770 m, whereas the similar estimate for M3 is 715 m with an adjustment of 35 m for land uplift (Table 2, Fig. 19).

#### **TPW-ELAs during the Lateglacial from c. 18,000 to 13,000 cal. yr BP**

A former meltwater channel linked to moraine M4 terminates in the  $L_4$  shoreline at 87 m a.s.l. about 20,000-18,000 cal. yr BP, whereas the M5-M7 moraines can be linked to a glaciofluvial fan which is abraded by the  $L_1$  sea level correlated to the Skarpsnes event (c. 14,500-14,000 cal. yr BP) (Figs. 4, 7 and 8).

Adjusted for land uplift, the lowering of the TPW-ELAs for the M4-M7 glaciers were 605, 590, 540 and 465 m, respectively (Table 2, Fig. 19).

#### **TPW-ELAs during the Younger Dryas (c. 13,000-11,500 cal. yr BP)**

A former meltwater channel linked to the ice-cored moraines M8 and M7 terminates in the  $L_0$ /Main shoreline formed during the Younger Dryas (12,800-11,500 cal. yr BP). Based on an AAR of 0.6, the estimated lowering of the TPW-ELA adjusted for land uplift is 390 m for the M8 glacier, whereas the similar estimate for the M9 glacier based on the modified 'Little Ice Age approach' (Dahl et al., 2002) is 380 m (Table 2, Fig. 19). This is comparable with estimates obtained from Øksfjordjøkulen, Svartfjelljøkulen and Langfjordjøkulen (Evans et al., 2002).

#### **TPW-ELAs during the early-Holocene (11,500 - 8800 cal. yr BP)**

The M-10, M11 and M12 glaciers are AMS radiocarbon dated on related lacustrine sediments in Aspvatnet to 10,400-10,300, 9800-9400 and 9300-8900 cal. yr BP, respectively. The estimated lowering of the TPW-ELA adjusted for land uplift is 300 m for the M10 glacier, 285 m for the M11 glacier and 280 m for the M12 glacier (Table 2, Fig. 19).

#### **TPW-ELAs during the mid-Holocene (8800-3800 cal. yr BP)**

After c. 8800 cal. yr BP, there was no input of glacier-derived meltwater sediments to Aspvatnet until 3800 cal. yr BP. Hence, the corresponding theoretical ELA must have been above c. 1200 m for the entire time span (Fig. 19).

#### **TPW-ELAs during the late Holocene (3800 - 0 cal. yr BP)**

After 3800 cal. yr BP, there has been a continuous input of glacier-meltwater derived sediments to Aspvatnet. To obtain a continuous TPW-ELA record

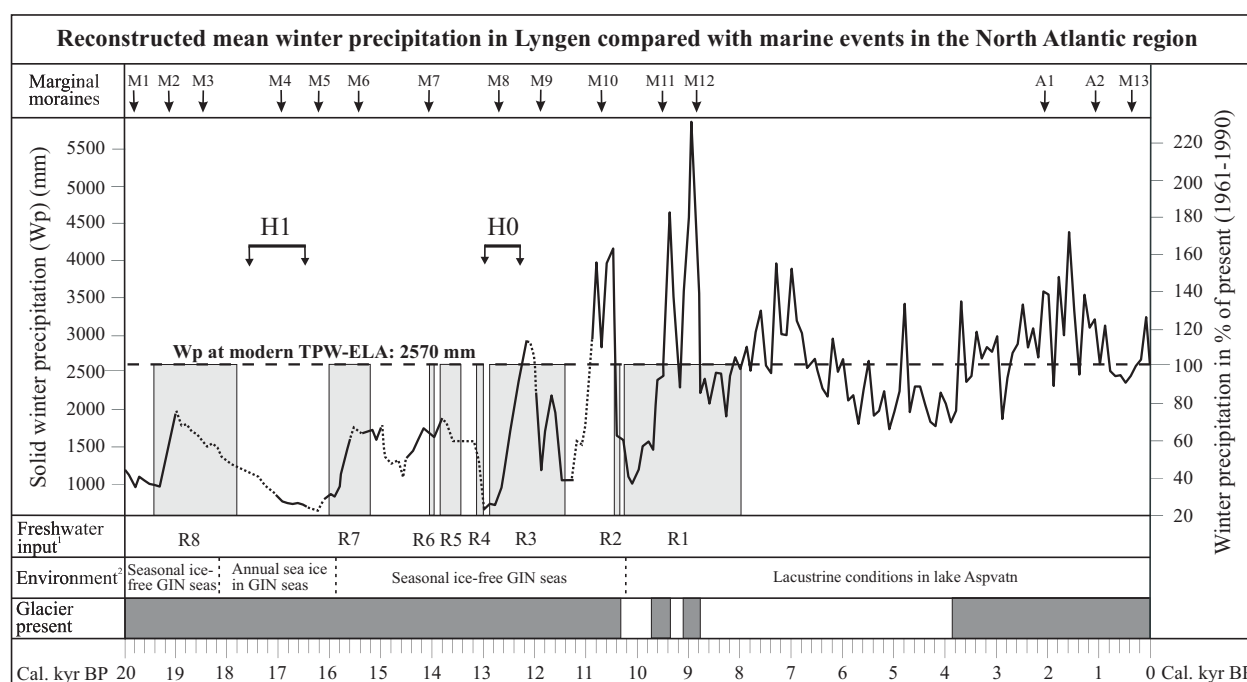
for this time span, a calibrated dry bulk density (DBD) curve (Fig. 17) has been used. Major glacier events are suggested to have taken place c. 1800 and 1200 cal. yr BP (A1 and A2 on Figure 19), but moraines from these episodes were probably erased by the maximum 'Little Ice Age' glacier c. AD 1910. Based on an AAR of 0.6, the TPW-ELA of the 'Little Ice Age' maximum M13 glacier was c. 120 m, whereas the suggested TPW-ELAs for A1 and A2 were lowered by about 80-40 m compared to modern values (Table 2, Fig. 19). From southern Lyngen the ELAs lowering during the 'Little Ice Age' (AD 1880-1910) were calculated to 100-250 m compared to modern values (Gellatly, 1988; Ballantyne, 1990).

### **Temperature reconstructions from northern Scandinavia**

Lateglacial temperature reconstructions are sparse from terrestrial sites in northern Scandinavia, as most of the land areas were covered by the continental Scandinavian Ice Sheet. An exception is the northern part of Andøya, where studies using sedimentological and biological proxies from the lakes Endletvatn, Øvre Ærásvatn and Nedre Ærásvatn have provided detailed information regarding the climatic development in the coastal regions of northern Scandinavia from c. 22,000  $^{14}\text{C}$  yr BP to the Younger Dryas/Holocene transition (Vorren et al., 1988; Alm, 1993). Hence, for the Lateglacial winter precipitation estimates in this study, a mean summer temperature reconstruction based on pollen from northern Andøya from c. 20,000 cal. yr BP to the Younger Dryas/Holocene transition c. 11,500 cal. yr BP has been used (Fig. 20) (Alm, op. cit.).

For the Holocene, several selected continuous temperature reconstructions based on biological proxies, like pine-tree limit fluctuations (Alm, 1993; Vorren and Alm, 1999; Vorren et al., 1999), pollen (Seppä and Birks, 2001; Korhola et al., 2002; Seppä and Birks, 2002; Seppä et al., 2002a) and chironomids (Korhola et al., 2002; Seppä et al., 2002b) are available from northern Scandinavia (Fig. 21). There are, however, major deviations between reconstructions based on pollen and chironomids from Finnmark and western parts of Finland. Based on the present knowledge it is difficult to evaluate which reconstruction is the most representative for Lyngen. Winter precipitation estimates using the different summer temperature reconstructions are therefore shown. The Holocene temperature reconstructions used in this study are adjusted for land uplift based on local sea-level curves or by inland extrapolation using shoreline diagrams from Troms and the Gulf of Bothnia.





**Figure 23** Variations in mean winter precipitation from 20,000 cal. yr BP to at present based on reconstructed TPW-ELAs from Strupskardet in Lyngen, Lateglacial temperature estimates from Andøya (Vorren *et al.*, 1988), early Holocene (11,500 to 10,500 cal. yr BP) temperature estimates from Ifjord (Seppä *et al.*, 2002a) and Holocene (10,500 cal. yr BP to at present) temperature estimates from Dalmutladdu (Bjune *et al.*, in press). The upper panel shows glacial events in Strupskardet (M1-M13), including the inferred glacial advances A1 and A2 lacking marginal moraines, Heinrich events are marked with H1 and H0, light-grey shaded areas indicate the timing of rerouting events of meltwater from the Laurentide Ice Sheet to the North Atlantic (Clark *et al.*, 2001)<sup>1</sup>, shifting environments in the Nordic sea are based on marine diatom reconstructions (Koc, 1993)<sup>2</sup>, and lower grey-shaded area indicates presence of glaciers in Strupskardet.

## Lateglacial and Holocene winter precipitation variations in Lyngen

The quantification of variations in Lateglacial and Holocene winter precipitation in Lyngen is based on equation 1, and follows the procedures outlined by Dahl and Nesje (1996).

The winter precipitation estimates from Strupskardet in Lyngen are mainly dependent on the accuracy of the temperature reconstructions and by the time resolution for the calculated former ELAs. Estimates from the Lateglacial reflect uncertainties regarding the length and timing of each event, and that only one temperature reconstruction with centennial time resolution has been available.

When Lenangsbreene were melted away from c. 8800 to 3800 cal. yr BP, the reconstructed winter precipitation curve reflects maximum values without glacierization in the catchment of Lenangsbreene. After 3800 cal. yr BP, the glacier magnitude and TPW-ELA of Lenangsbreene is suggested to be reflected by a calibrated DBD curve with a time resolution of c. 30 years. Hence, the quantified variations in winter precipitation have increased accuracy for this time span (Fig. 22).

Lenangsbreene consist of two nearly coalesced cirque glaciers which receive a large amount of windblown snow compared to the regional mean in this part of Lyngen. Hence, the ELA of Lenangsbreene reflects a local TPW-ELA. As

Lenangsbreene is part of a larger ice field, including glaciers in most aspects, however, the combined TPW-ELAs of these glaciers may be regarded as representative for a regional TPW-ELA in Lyngen (Fig. 5). Based on the inferred 'Little Ice Age' and Younger Dryas moraines in Figure 5, this is even more likely when these glaciers are in a more advanced position than at present.

As demonstrated by the reconstructed moraines in Figure 5, the relative importance of glaciers with different aspect was more-or less similar to the present during the 'Little Ice Age' and the Younger Dryas. If representative for all the reconstructed former glaciers in Strupskardet, this implies that the reconstructed Lateglacial and Holocene winter precipitation curves in Figures 22 and 23 are representative for the former regional winter climate in Lyngen.

## Comparison with Holocene glacier records in northern Scandinavia

The Holocene glacial history obtained in this study deviates in some periods with that suggested for northern Sweden (Karlén, 1973; Karlén and Denton, 1975; Karlén, 1976) and for northern Norway (Griffey and Worsley, 1978; Ballantyne, 1990; Winkler *et al.*, 2003). However, the inferred 'Little Ice Age' maximum c. A.D. 1900 at Lenangsbreene and the lacustrine record from Aspvatnet, indicating that the

glacier was almost as large c. A.D. 1800, fit well with previous studies from northern Scandinavia. At some sites at Svartisen, Okstindan, Saltfjellet and Kebnekaise, the late-Holocene glacier maximum may have occurred earlier, with advances dated at A.D. 1750, 1810, 1850 and 1890 (Karlén, 1976).

Evidence of late-Holocene glacier advances between 3000 and 2000 cal. yr BP beyond the 'Little Ice Age' maximum have been reported at some glaciers based on lichenometry in northern Scandinavia (Karlén, 1976; Ballantyne, 1990), and supported by radiocarbon-dated palaeosols at Okstindan (Griffey and Worsley, 1978). These advances may correspond to the inferred glacier advances A1 and A2 somewhat smaller than the 'Little Ice Age' maximum at Lenangsbreen (Figs. 19 and 23). However, the lichen dated 'old' glacier advances in northern Sweden and Lyngen are well beyond the reliable range of modern growth curves in Norway (e.g. Innes, 1985; Erikstad and Sollid, 1986; Matthews, 1994; Winkler et al., 2003).

During the early-Holocene, the M10, M11 and M12 glacier advances took place c. 10,400-10,300, 9800-9400 and 9300-8900 cal. yr BP in Strupskardet. Without exact age estimates, this is in accordance with several moraines of suggested early-Holocene or Younger Dryas age reported by Ballantyne (1990) in Lyngen.

Corner (1980) reported glacial events related to the late deglaciation of the Scandinavian Ice Sheet dated to c. 9900-9800 ± 150, 9600-9500 ± 150 and 9400 ± 150 <sup>14</sup>C yr BP in Storfjorden east of Lyngen. The rapid retreat following the Younger Dryas readvance in Norway has been linked to glacier dynamics due to calving in the fjords (e.g. Holtedahl, 1975; Andersen, 1980; Andersen et al., 1995). Commonly, ice-frontal deposits are located on rock thresholds where the fjords become shallower, and/or where the valleys/fjords are relatively narrow. Because of rapid calving, steep and dynamically unstable glacier profiles were established. Hence, when the glacier front became grounded and more dynamically stable, still partly active glaciers had minor advances in response to the steep profiles. The reported glacial events in Storfjorden east of Lyngen are therefore not regarded as climatic induced advances.

### **Comparison with glacier records from southern Norway**

The combined Holocene glacier records from southern Norway reflect complex pattern(s) like:

- The degree of continentality with different responses to changes in summer temperature and winter precipitation.
- Morphological differences related to altitude, aspect, gradients and magnitude.
- Different atmospheric circulation patterns

during for instance the 'Little Ice Age'.

Regarding major glacier events, however, it still makes sense to compare southern Norway with northern Scandinavia.

In southern Norway there is evidence of glacial advances during the 'Little Ice Age' c. A.D. 1750, 1780 – 1820, 1830, 1850, 1870-1890 and 1930 at different glaciers existing in continental to maritime climate regimes (Andersen and Sollid, 1971; Tvede and Liestøl, 1977; Hole and Sollid, 1979; Erikstad and Sollid, 1986; Grove, 1988; Dahl and Nesje, 1996; Nesje et al., 2001). In Lyngen the LIA glacier maximum is suggested to be c. A.D. 1910 (Ballantyne, 1990; this study), and similar late LIA glacier advances are found along the coast of western Norway (e.g. Tvede and Liestøl, 1977; paper 3). Glacier advances beyond the LIA glacier maximum c. 3000 to 2000 cal. yr BP has so far only been reported from Okstindan (Griffey and Worsley, 1978), but have now also been found at some outlet glaciers at northern Folgefonna (paper 3).

Most studied glaciers in Scandinavia seem to have been completely melted away (or very small) during at least parts of the early- to mid-Holocene climate optimum, and the timing for the onset of neoglaciation is rather synchronous from 5800 to 2100 cal. yr BP based on factors like continentality, morphological differences etc. For instance, both Hardangerjøkulen in central southern Norway and Lenangsbreen have existed continuously since c. 3800 cal. yr BP (e.g. Dahl and Nesje, 1994; this study).

During the mid-Holocene thermal optimum, however, there are several glacier advances like the Finse Event ('8.2 ka event') at both Hardangerjøkulen and Jostedalbreen that apparently are lacking in Lyngen (e.g. Dahl and Nesje, 1996; Nesje et al., 2001). Prior to the Holocene thermal optimum several glacial expansion episodes at the maritime to semi-continental glaciers Folgefonna, Hardangerjøkulen and Jostedalbreen have been recorded, with glacier advances c. 11,300 – 11,100, 10,600 – 10,500, 10,100-9900, 9800-9700 and 8300 – 8100 cal. yr BP. (Dahl and Nesje, 1996; Nesje et al., 2001, paper 3). However, some of these events may be related to the M10, M11 and M12 glaciers in Strupskardet, dated at 10,400-10,300, 9800-9400 and 9300-8900 cal. yr BP, respectively.

### **Comparison with marine climate archives in the North Atlantic region**

To put the inferred Lateglacial and Holocene climate from Lyngen in a North Atlantic context, the winter-precipitation record from Strupskardet has been compared with the timing of Heinrich events (Clark et al., 2002), melt-water rerouting events from the Laurentide ice sheet (Clark et al., 2001) and the sea-ice record based on marine diatoms (Koc et al., 1993) (Fig. 23). The mean Lateglacial winter precipitation

in Lyngen was c. 50–60 % compared to at present and seasonal sources of precipitation is suggested to have existed. Open water in the Norwegian Sea, however, assumes production of North Atlantic Deep Water (NADW), and according to Dokken and Jansen (1999) this may have taken place along the western coast of Norway because of brine formation during Weichselian stadials.

Being simultaneous with Heinrich events 1 (H1) and 0 (H0) (Clark *et al.*, 2002), the two driest periods during the Lateglacial in Lyngen had mean winter precipitation values below 30 % compared to at present (Fig. 23). During H1, the Nordic Seas experienced an annual sea-ice cover based on marine diatoms (Koc *et al.*, 1993), whereas the much shorter H0, had seasonally open water. Hence, based on the winter precipitation curve from Lyngen, the atmospheric circulation responded to the SST lowering associated with H1 and H0.

Except for R4 and R2, no periods with freshwater input to the North Atlantic (e.g. Clark *et al.*, 2001) resulted in a decreased mean winter precipitation in Lyngen. On the contrary, episodes R8, R7, R6 and R5 had relatively high mean values, whereas significant maxima occurred during the Lateglacial in R3 and during the early-Holocene in R1 (Fig. 23).

## Main Conclusions

Based on the presented lacustrine and morphostratigraphical evidence from Lyngen in Troms, the following conclusions of local and regional importance are suggested:

1. Thirteen marginal moraines have been mapped in front of Lenangsbreene in Strupskardet, Lyngen. Moraines M1–M3 and M8–M9 are suggested to be relict ice-cored moraines, whereas moraines M4–M7 and M10–M13 are push and melt out moraines.

2. Based on a combination of relict periglacial features, former shorelines and related glacier-meltwater induced channels, lichenometry and AMS radiocarbon dated lacustrine sediments, a chronological framework spanning from 20,000 cal. yr BP to the present has been established.

3. The proglacial lake Aspvatnet was isolated from the sea c. 10,300 cal. yr BP. Based on LOI at 550°C, LOI at 950°C, magnetic susceptibility (MS), water content, wet- and dry bulk density (DBD), anhysteretic remanent magnetisation (ARM) and saturation isothermal magnetisation (SIRM), the lacustrine phase was divided into the following modes: periods with input of glacier meltwater in a brackish environment (mode GB-1), periods with input of glacier meltwater in a lacustrine environment (modes G-1A/G-1B, G-2 and G-3), periods with a lacustrine environment without glaciers in the catchment (modes L-1 to L-6) and river floods in a lacustrine environment (modes F-1 to F-3). There is in general a good agreement between physical sediment parameters and magnetic

parameters. The age-depth model is based on ten AMS radiocarbon dates on terrestrial plant macrofossils, marine shells and bulk gyttja samples.

4. Dry bulk density (DBD) is inferred to be representative for glacier variations during the last 3800 cal. yr BP, and has been calibrated to reflect a continuous TPW-ELA curve for this period. The correlation between dry bulk density and a combination of medium and fine silt is 0.89. Episodes when the two parameters 'sorting' and 'mean' are in strong and abrupt anti-phase, are suggested to represent rapid events like river floods with a paraglacial signature.

5. Using a combination of an AAR of 0.6 and an adapted 'Little Ice Age ratio' approach, thirteen former TPW-ELAs adjusted for land uplift have been estimated. The TPW-ELA in Strupskardet has varied from a maximum of c. 1200 m (>240 m above the present TPW-ELA) to a lowest altitude of c. 100 m (c. 840 m below the present TPW-ELA). As Lenangsbreene are part of a larger ice field, with glaciers in all aspects, the TPW-ELAs from this glacier are suggested to be representative for the regional TPW-ELA in Lyngen.

6. Mean winter precipitation has been quantified based on equation 1 (the 'Liestøl equation'). In Lyngen, the reconstructed Lateglacial and Holocene mean annual winter precipitation as snow varied from more than 5000 mm water equivalents to 500–600 mm, or from more than 200 to 25 per cent (the present mean winter precipitation (1 October–30 April) is equal to 100%). The mean winter precipitation during the Holocene was about twice the similar mean estimate for the Lateglacial.

7. The 'Little Ice Age' glacier maximum was c. A.D. 1900 in Lyngen. In general the reconstructed Holocene glacier record from Lyngen fits rather well with previous published records from northern Scandinavia. The glacier advances c. 3000 to 2000 cal. yr BP beyond the LIA maximum at Okstindan, are suggested to have been somewhat smaller in Lyngen. The major glacier events during the early and late Holocene in southern Norway seem to be more-or-less synchronous with northern Scandinavia, whereas the mid Holocene glacier advances (e.g. the Finse Event/'8.2 ka event') in southern Norway apparently are missing in Lyngen.

8. The timing of H1 and H0 during the Lateglacial occurred simultaneously with the two driest periods with inferred formation of talus-derived rock glaciers in Lyngen. However, only the first and rather long dry period coincided with an annual sea-ice cover in the Nordic seas (Koc *et al.*, 1993). Among eight episodes with freshwater input to the North Atlantic (e.g. Clark *et al.*, 2001), only two may have had an impact on the mean winter precipitation in Lyngen.

9. Seasonal open water with brine formation is suggested to have existed along the western coast of Norway during Weichselian stadials. However, as the average winter precipitation during H1 and H0



was c. 50 per cent of the Lateglacial mean (and 25 per cent of the present values), it implies that the atmospheric circulation may have responded to the reduced SST during these events.

10. The quantification of mean winter precipitation from Lyngen is depend on the reconstructed summer temperatures and the time resolution and amplitude of former TPW-ELAs. The inferred uncertainties from the Lateglacial are mainly related to the length and timing of each event, and that there only exists one temperature reconstruction from this period. The apparent discrepancies between Holocene temperature records from northern Scandinavia represent a problem for precise estimates of winter precipitation during this period.

11. Lyngen has the potential to yield terrestrial quantitative palaeoclimatic reconstructions from 20,000 cal. yr BP to present that can be used to investigate ocean-atmosphere dynamics.

### Acknowledgements

This is a contribution to NORPEC – a strategic university project at the University of Bergen coordinated by Professor H. John Birks. Thanks to Kristina Pedersen, Joachim Riis Simonsen, Åsmund Bakke, Lars-Ivar Folgerø, Bjørn Kvisvik and Erlend Paasche for assistance during the fieldwork. Thanks to Øyvind Lie for valuable comments to the manuskript.

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