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

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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 = ~ 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 ~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 glacigenic 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.*, 200a; 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

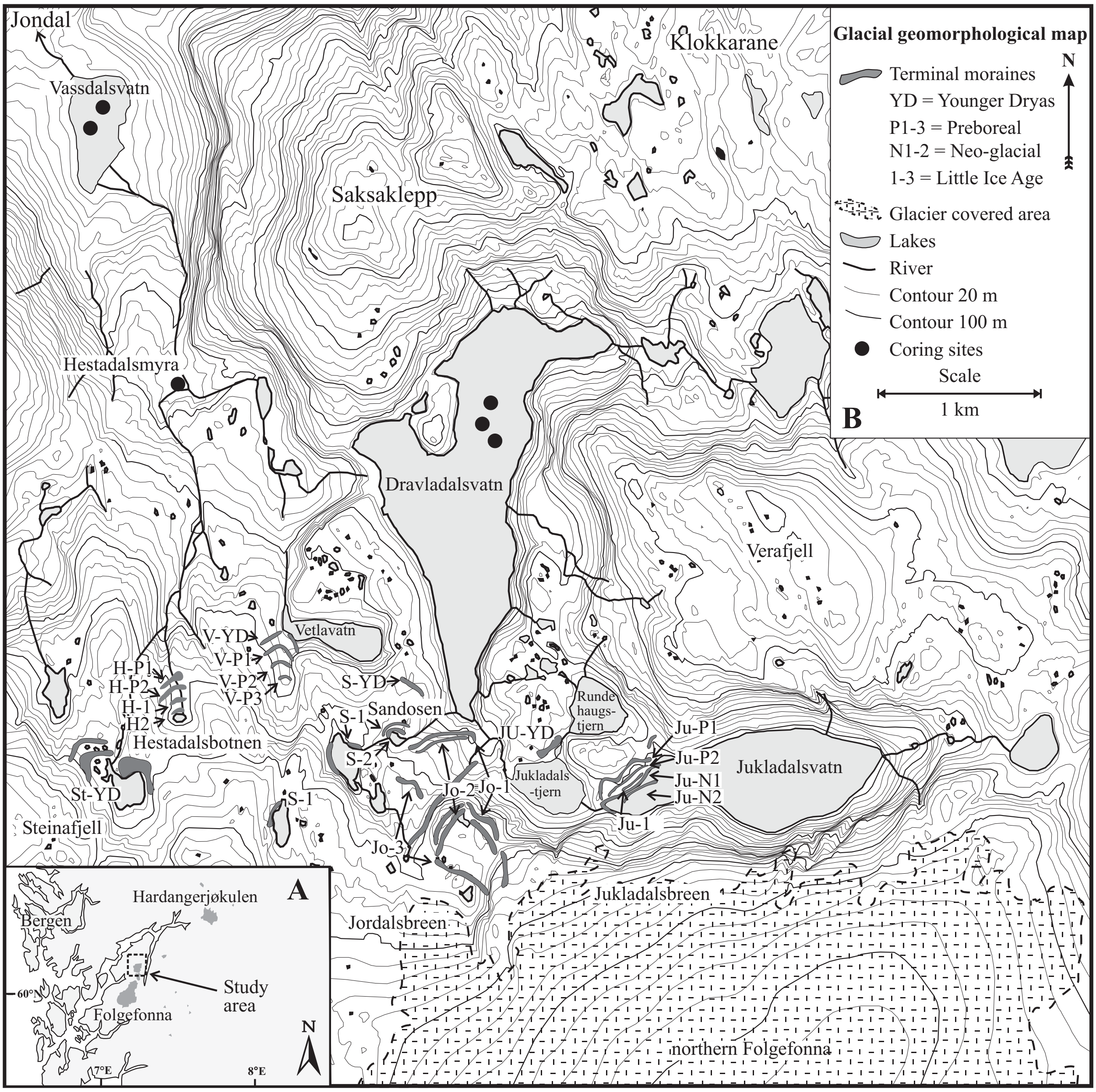

Figure 1 (A) Map over southern Norway including glaciers (Folgefonna and Hardangerjøkulen) adapted from Østrem *et al.* (1988). (B) Detailed map over the moraine systems in the catchment north of the plateau glacier northern Folgefonna. Grey areas are terminal moraines mapped in the catchment. Black 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^2 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, Dettebrea and Juklavassbreen. About 12 km^2 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 (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), this gives a mean Ts 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^2 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^2 (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^2 (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^3) (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' (µm) 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.

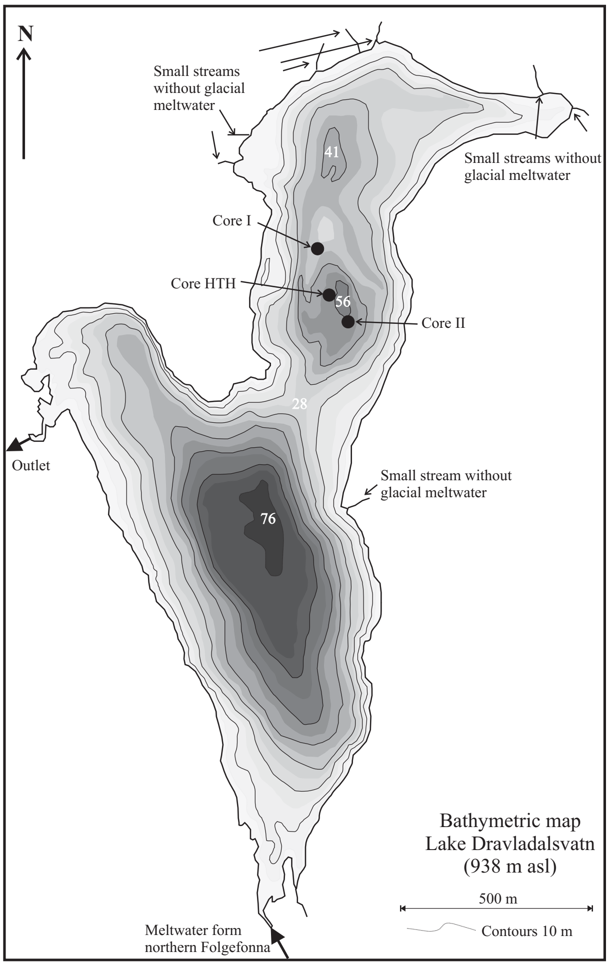

Figure 2 Bathymetric map of Dravladalsvatn showing the location of the retrieved piston cores and the short gravity core (HTH). Note the bedrock threshold separating the two main basins with a 28 m deep sill. Inflow and outflow of glacial meltwater from northern Folgefonna are both in the western main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

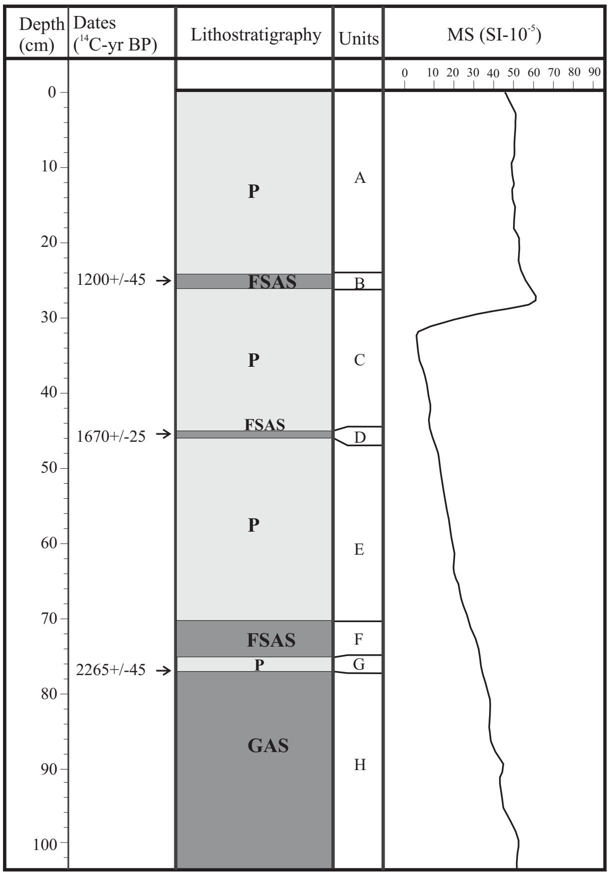

Figure 3 Lithostratigraphy and MS in a core retrieved from the peat bog from Hestadalsmyra. Units B, D and F represent periods with glacial input to the peat bog (P = peat, FSAS = fine sand and silt, GAS = gravel and sand). The MS signal shows peculiar patterns, gradually decreasing up to unit B were 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±45 ^{14}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±25 ^{14}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±45 ^{14}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±70 ^{14}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±40 ^{14}C yr BP (Poz-3177) and of 5530±40 ^{14}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±45 ^{14}C yr BP (TUa-3628A). The upper part of the unit was radiocarbon dated at 2000±40 ^{14}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±30 ^{14}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±70 ^{14}C yr BP (TUa-3632A) and in the upper part 5050±70 ^{14}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±35 ^{14}C yr BP (Poz-3179). Unit A consisted of grey clay and silt with an age in the lower part of 3215±60 ^{14}C yr BP (TUa-3631A). At 78 cm a

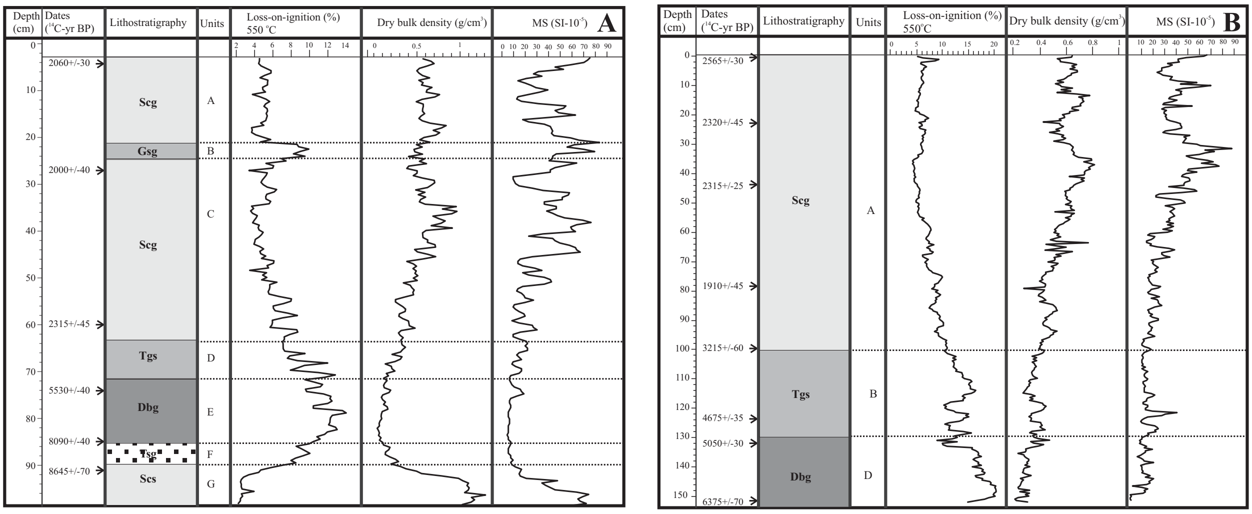

Figure 4 (A) Lithostratigraphy, radiocarbon dates (^{14}C-yr BP), loss-on-ignition (% loss at 550 ^{o}C), DBD and MS measured at 0.5 cm interval in Dravladalsvatn core I (Scg = grey silt / clay, Dbg = dark brown gyttja, Tst = transition from silt towards brown gyttja, Scs = silty clay with stones, Tgs = transition from gyttja to silt / clay, Gsg = grey silty gyttja). (B) Lithostratigraphy, radiocarbon dates (^{14}C-yr BP), loss-on-ignition (% loss at 550 ^{o}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}C yr BP (TUa-3630). In the upper part of unit A, three inverted radiocarbon dates were obtained, yielding 2315±25 ^{14}C yr BP (Poz-3798), 2320±45 ^{14}C yr BP (TUa-3640A) and 2565±30 ^{14}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-frequent 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}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}C yr BP (Beta-102935) and 4270±80 ^{14}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}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}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}C yr BP (Beta-102930). The LOI record shows an abrupt change from the lower minerogenic section (unit G) into unit F. The MS

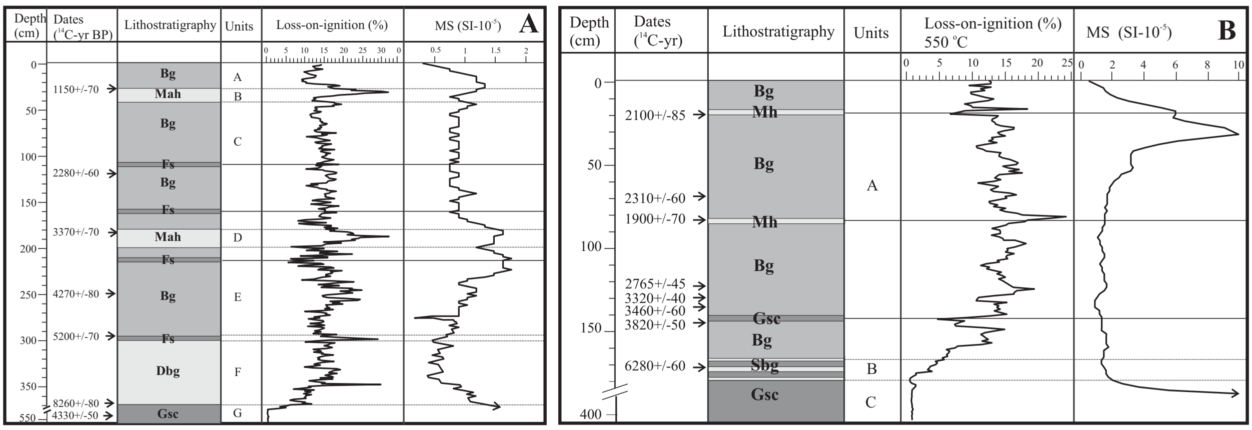

Figure 5 (A) Lithostratigraphy, radiocarbon dates (^{14}C-yr BP), loss-on-ignition (% loss at 550 ^{o}C) and MS measured at 2-cm intervals for Vassdalsvatn core I (Bg = brown gyttja, Mah = macrofossil and humus, Fs = fine sand and silt, Dbg = dark brown gyttja). (B) Lithostratigraphy, radiocarbon dates (^{14}C-yr BP), loss-on-ignition (% loss at 550 ^{o}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).

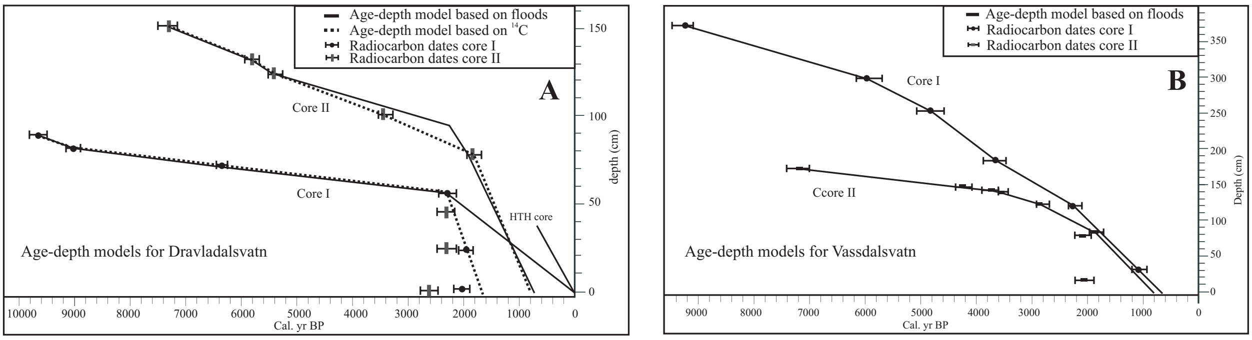

Figure 6 (A) Age-depth models for the three cores retrieved in Lake Dravladalsvatn. The solid line shows the linear interpolation model based on the radiocarbon dates in the each core. Dotted line shows the age-depth model correlated against the lithostratigraphy in Vassdalsvatn and the peat in Hestadalsmyra. 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±60 ^{14}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±50 ^{14}C yr BP (UtC-6694) below and 3460±60 ^{14}C yr BP (UtC-6693) above. From this unit several radiocarbon dates have been obtained; 3320±40 ^{14}C yr BP (UtC-6692), 2765±45 ^{14}C yr BP (UtC-6691), 1900±70 ^{14}C yr BP (T-13608), 2310±60 ^{14}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 ± 1 sigma	Intercept (cal. yr BP)	±1 sigma (cal. yr BP)	±2. sigma (cal. yr BP)
Vassdalsvatn	**Core I**						
Vassdalsvatn I	Beta-102930	28-31	Gyttja	1150±70	1060	1170-970	1185-930
Vassdalsvatn I	Beta-102931	117-120	Gyttja	2280±60	2240	2350-2160	2360-2120
Vassdalsvatn I	Beta-102932	182-185	Gyttja	3370±70	3645	3690-3480	3830-3460
Vassdalsvatn I	Beta-102933	250-253	Gyttja	4270±80	4810	4965-4650	5045-4570
Vassdalsvatn I	Beta-102934	295-298	Gyttja	5200±70	5960	6170-5890	6170-5750
Vassdalsvatn I	Beta-102935	368-372	Gyttja	8260±80	9245	9415-9130	9430-9060
Vassdalsvatn I	Beta-102936	525-535	Gyttja	4330±50	4910	4965-4840	4990-4830
Vassdalsvatn	**Core II**						
Vassdalsvatn II	T- 13607	19	Gyttja	2100±85	2050	2295-1970	2210-1890
Vassdalsvatn II	T-13608	83-84	Gyttja	1900±70	1840	1920-1735	1990-1690
Vassdalsvatn II	T- 13788A	77-79	Gyttja	2310±60	2080	2360-2160	2470-2150
Vassdalsvatn II	UtC-6691	123	Gyttja	2765±45	2860	2920-2785	2950-2775
Vassdalsvatn II	UtC-6692	138	Gyttja	3319±40	3550	3630-3475	3640-3465
Vassdalsvatn II	UtC-6693	142	Gyttja	3460±60	3720	3830-3640	3870-3570
Vassdalsvatn II	UtC-6694	147	Gyttja	3820±50	4250	4345-4100	4410-4090
Vassdalsvatn II	UtC-6695	171	Gyttja	6280±60	7205	7270-7030	7405-7005
Hestadalsmyra							
Hestadalsmyra	T-13600	26-28	Humus	1200±45	1120	1255-995	1280-965
Hestadalsmyra	T-13601	45,5-47,5	Humus	1670±65	1560	1690-1515	1715-1410
Hestadalsmyra	T-13602	75-77	Humus	2265±60	2250	2345-2160	2355-2145
Dravladalsvatnet	**Core I**						
Dravladalsvatn I	Poz-3175	1	Macro fossil	2060±30	2030	2060-1990	2115-1950
Dravladalsvatn I	TUa-3627A	24	Gyttja	2000±40	1955	1990-1920	2045-1865
Dravladalsvatn I	TUa-3628A	57	Macro fossil	2315±45	2310	2355-2305	2465-2155
Dravladalsvatn I	Poz-3176	72	Macro fossil	5530±40	6340	6395-6290	6405-6280
Dravladalsvatn I	Poz-3177	82	Gyttja	8090±40	9055	9220-9000	9130-8980
Dravladalsvatn I	TUa-3629A	88	Gyttja	8645±70	9660	9690-9540	9800-9520
Dravladalsvatnet	**Core II**						
Dravladalsvatn II	Poz-3178	1	Gyttja	2565±30	2620	2750-2550	2755-2490
Dravladalsvatn II	TUa-3640A	24	Gyttja	2320±45	2310	2360-2180	2465-2155
Dravladalsvatn II	Poz-3198	45	Gyttja	2315±25	2330	2350-2330	2355-2310
Dravladalsvatn II	TUa-3630	78	Macro fossil	1910±45	1840	1910-1745	1950-1725
Dravladalsvatn II	TUa-3631A	100	Gyttja	3215±60	3450	3475-3360	3575-3330
Dravladalsvatn II	Poz-3179	124	Gyttja	4675±35	5390	5465-5320	5470-5310
Dravladalsvatn II	Poz-3256	132	Gyttja	5050±30	5810	5890-5805	5900-5725
Dravladalsvatn II	TUa-3632A	151	Gyttja	6375±70	7305	7415-7250	7430-7180

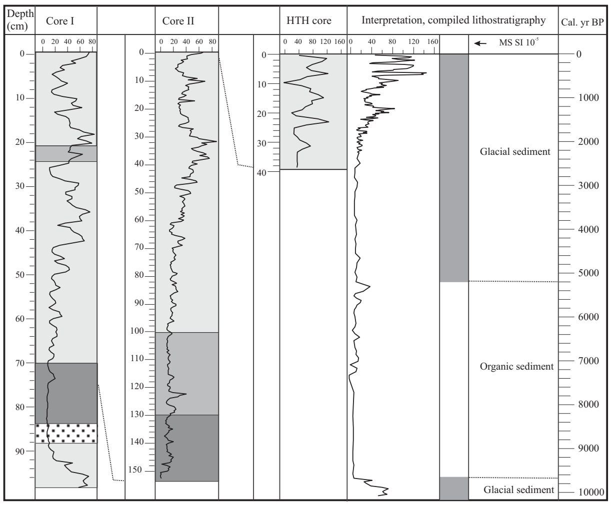

Figure 7 Compiled lithostratigraphy from Dravladalsvatn based on the three cores retrieved from the eastern distal basin. Correlation between the cores is done by using the age-depth models and the MS records.

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

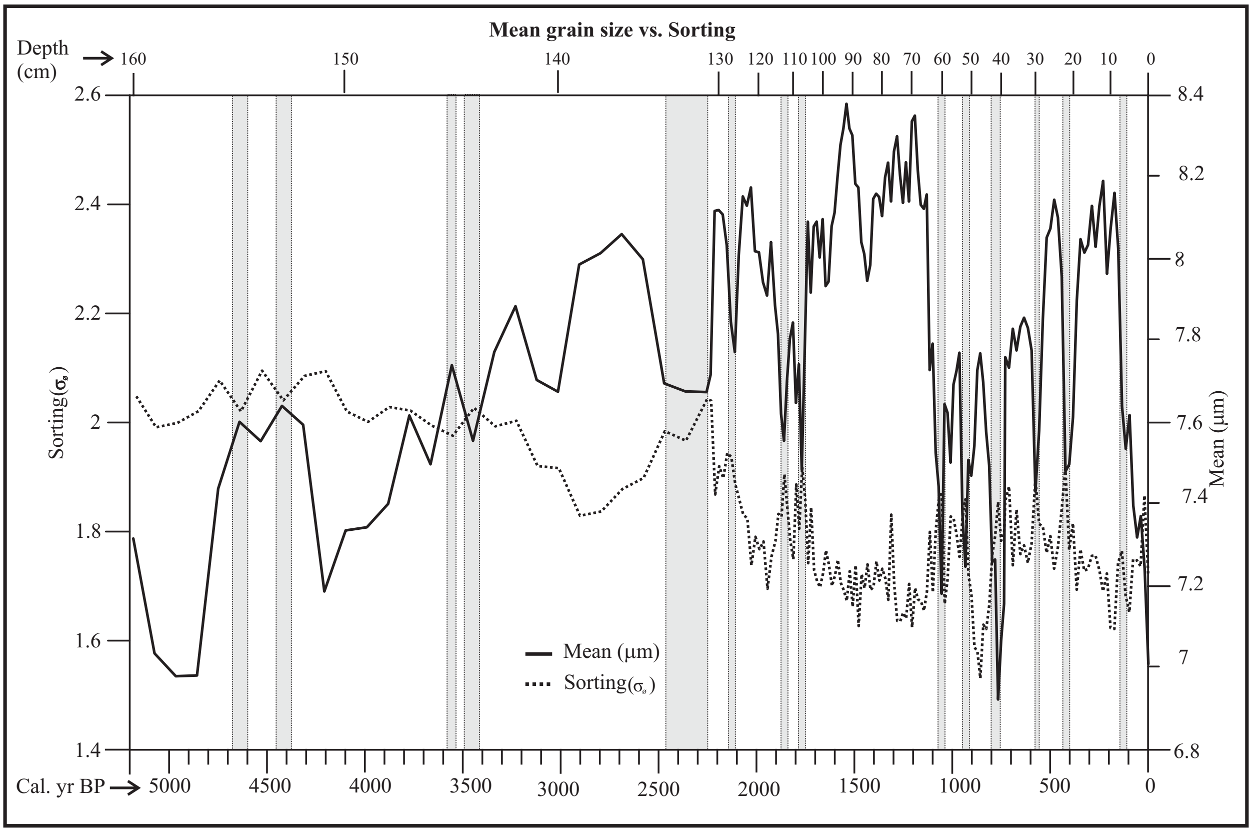

Figure 8 Mean grain-size plotted against sorting (standard deviation in a sample). The upper axis shows depth (cm) in the compiled stratigraphy and the lower axis shows calendar year before present. As seen from the depth scale, there is a notable change in sedimentation rate around 130 cm. Higher 'sorting'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

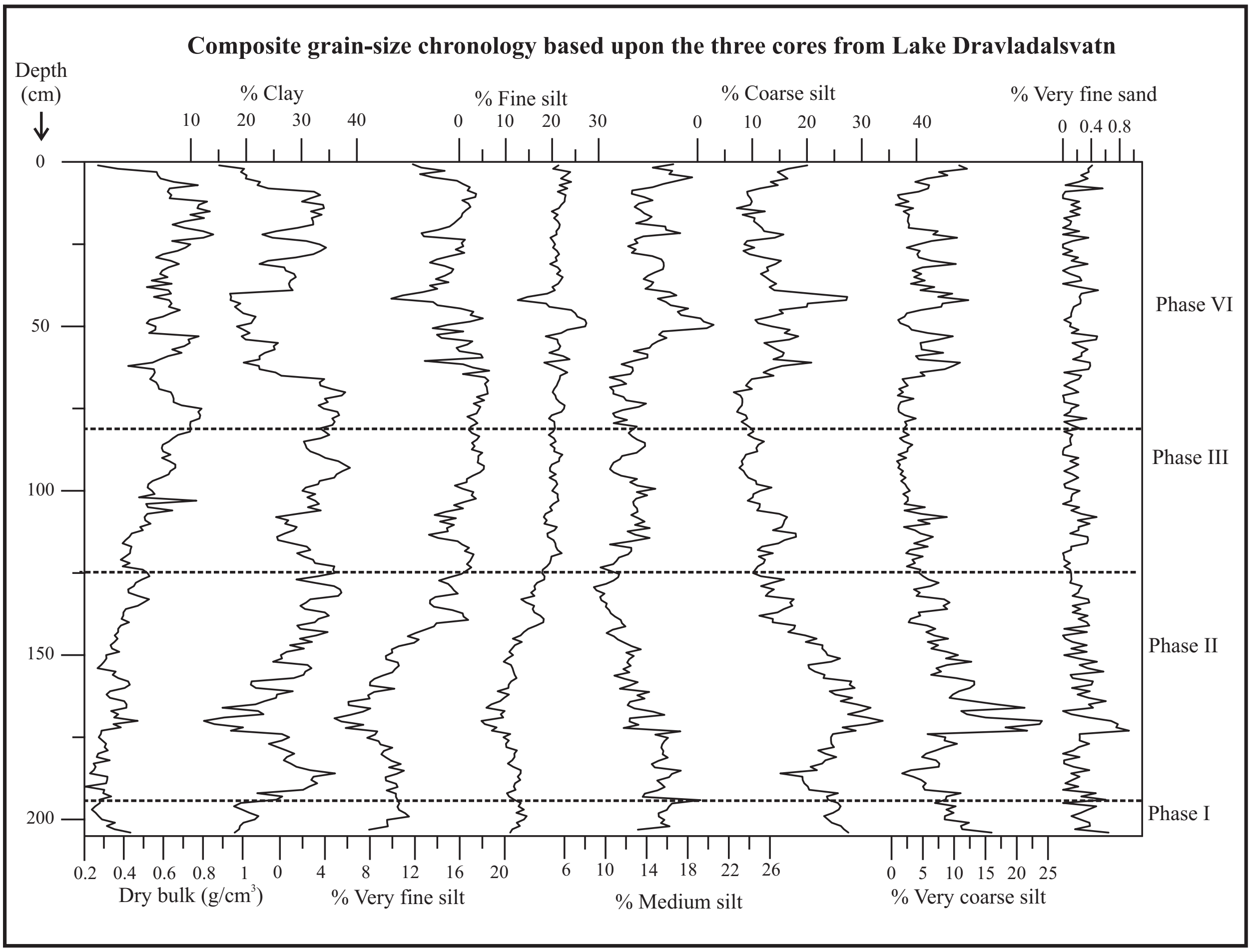

Figure 9 Bulk density, clay, very fine silt, fine silt, medium silt, coarse silt, very coarse silt and very fine sand. The compiled lithostratigraphy is divided into four phases based on the presented parameters were they indicate notable changes in sedimentation environment.

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^{-1}) (Figure 7).

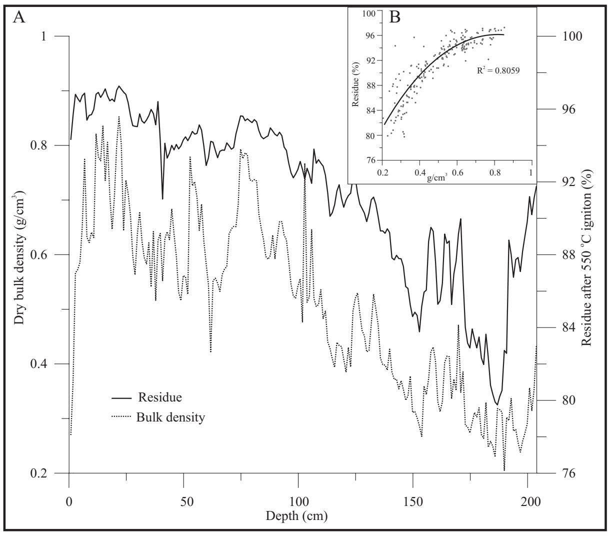

Figure 10 (A) Residue after 550°C ignition in % giving the minerogenic proportion after the organic content is removed. This parameter has traditionally been used as an indicator for inorganic sedimentation. Dotted line shows DBD. (B) Regression between residue (%) and DBD, showing a close relationship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

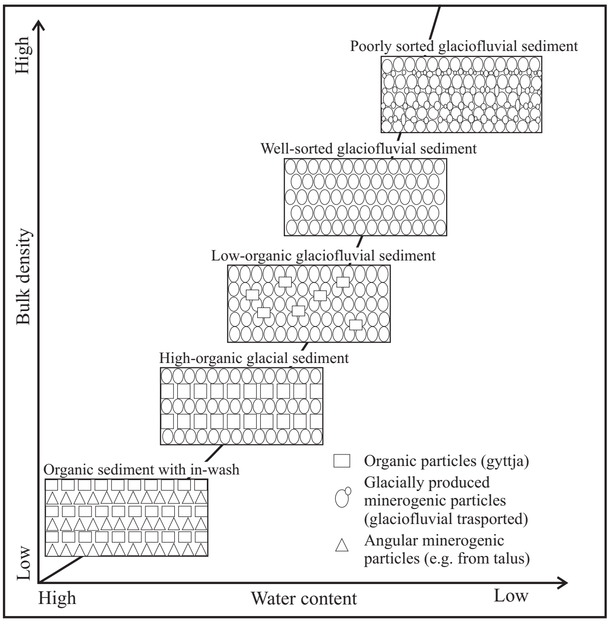

Figure 11 Schematic figure explaining the relationship between bulk density and water content related to type of sediment. Angular minerogenic particles give higher porosity than rounded glacial-meltwater derived minerogenic particles. Lowest bulk density values are obtained from sediments dominated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

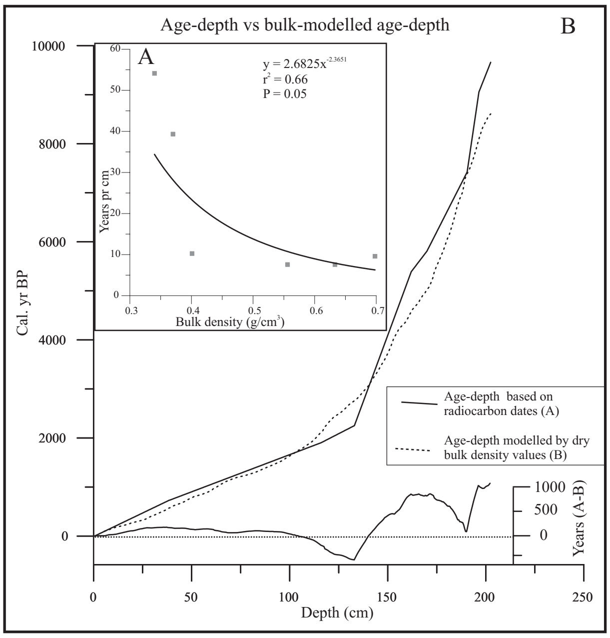

Figure 12 (A) Regression between bulk density values and time resolution (years pr centimetre). The coefficient of determination (R-squared) shows high predictability due to sedimentation rate through the bulk values. (B) Modelled age-depth based on average bulk values between the radiocarbon dates in 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.

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 glacigenic 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 complement Dravladalsvatn and for obtaining additional information regarding major flooding and glacier events in the catchment. The dicrepancey 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).

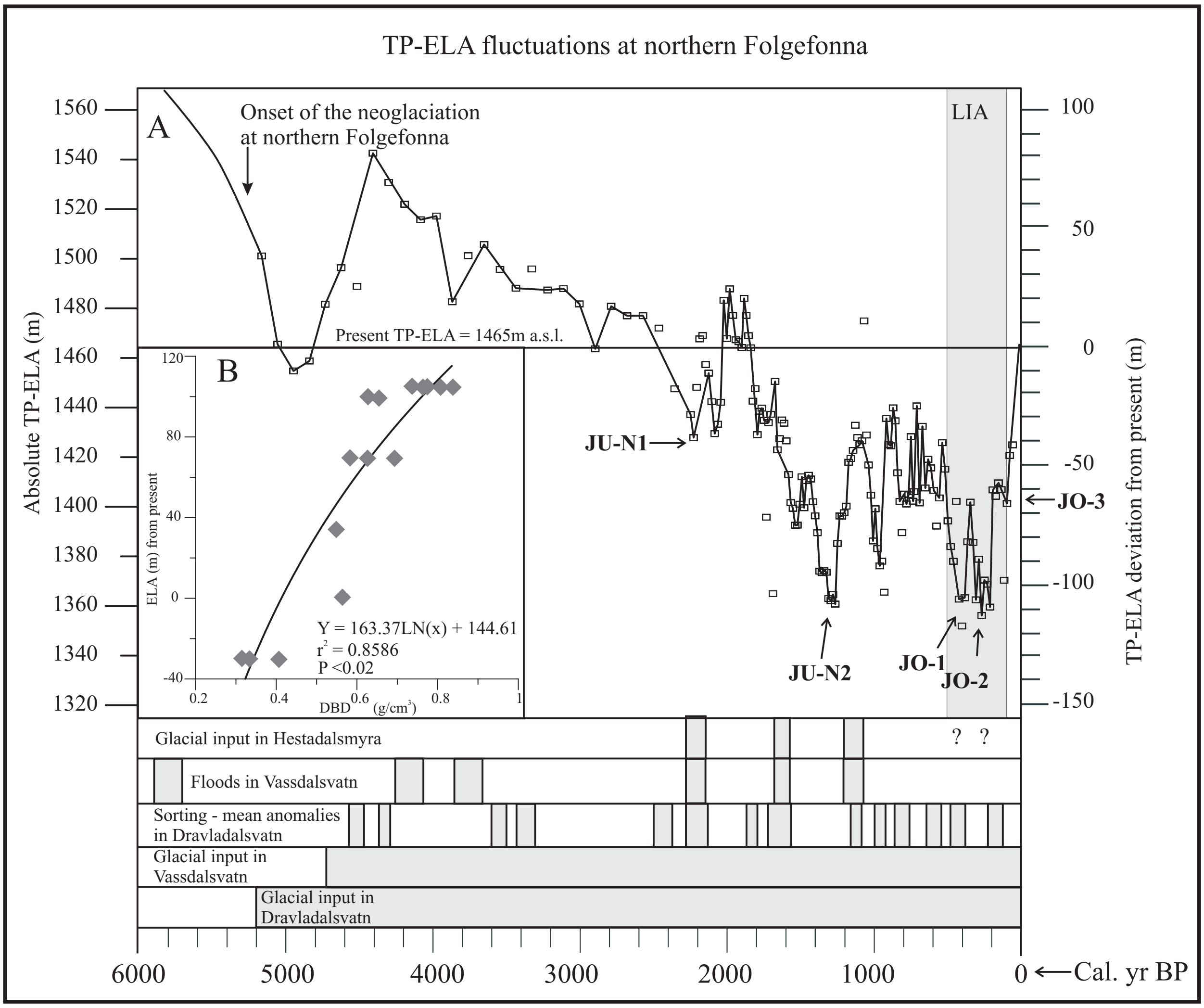

Figure 13 (A) TP-ELA curve for Northern Folgefonna based on the bulk density record. The lower part of the figure is a summary of sources for validating the ELA record; the peat bog in Hestadalsmyra, floods in Vassdalsvatn, sorting anomalies in Lake Dravladalsvatn, glacial input in Vassdalsvatn and record 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 glacigenic 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 µm), 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:

$$B_n = 444 + 2.16 * P - 54 * T_3,$$

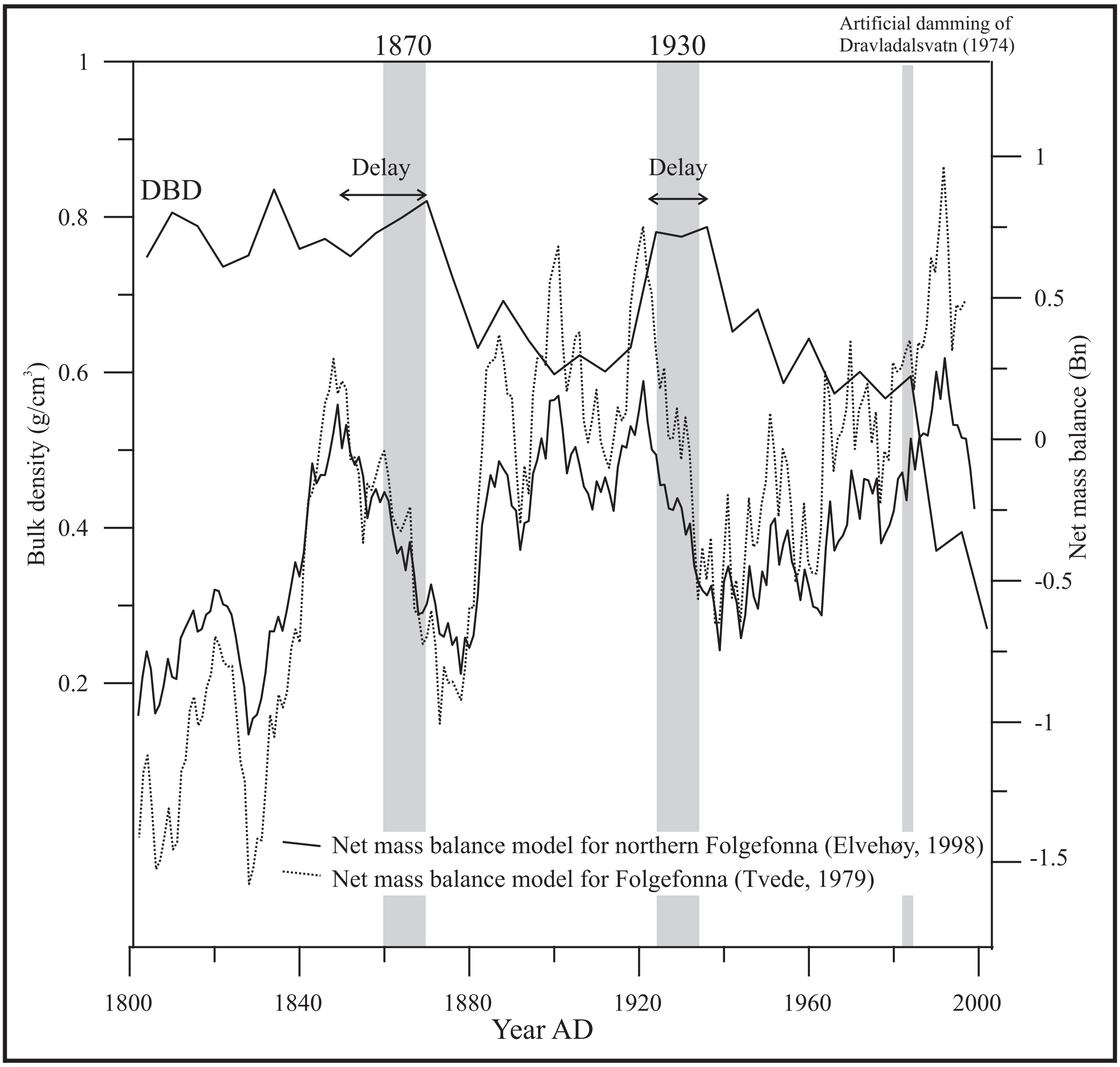

Figure 14 Dry bulk density (DBD) compared to two different glacier net mass-balance models for the Folgefonna and the Northern Folgefonna glaciers. The dotted line shows a model developed by Tvede (1979) for the southern part of the Folgefonna glacier, whereas the black line was produced for northern 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 Metrological 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 ~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.

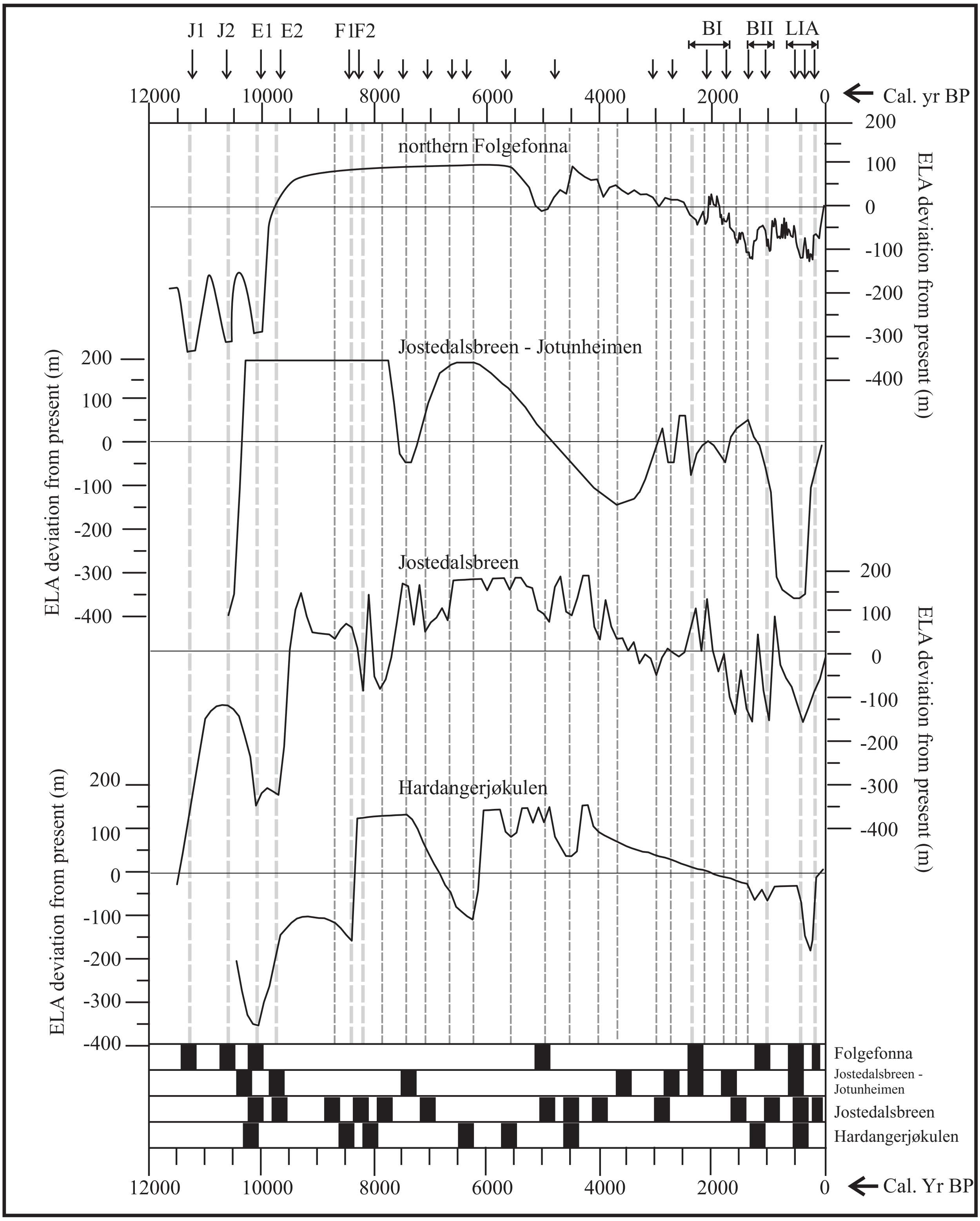

Figure 15 Compilation of some selected ELA reconstructions from southern Norway throughout the Holocene period (Paper 3 and 2, Matthews and Karlén, 1992, Nesje *et al.*, 2001, Dahl and Nesje, 1996). Upper panel with arrows shows glacial events (abrupt events of decadal to millennial duration) derived 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 ^{14}C yr BP and 1250-1000 ^{14}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 form 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 were 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.

References

Arnaud, F., Lignier, V., Revel, M., Desmet, M., Beck, C., Pourchet, M., Charlet, F., Trentesaux, A., and Tribovillard, N. 2002: Flood and earthquake disturbance of ^{210}Pb geochronology (Lake Antern, NW Alps). *Terra Nova* B072, 1-8.

Askvik, H. 1995. Oversikt over Norges Prekambriske og Paleozoiske berggrunn, pp. 1-75. Geologisk Institutt, Bergen.

Bakke, J. 1999. "Rekonstruksjon av bre- og klimavariasjonar på Nordre Folgefonna med kringliggjande botnar." Unpublished Thesis, Department of Geography, University of Bergen.

Ballantyne, C., and Benn, D. I. 1994: Paraglacial Slope Adjustment and Resedimentation Following Recent Glacier Retreat, Fåbergstølsdalen, Norway. *Arctic and Alpine Research* 26, 255 / 269.

Ballantyne, C. K. 2002: Paraglacial geomorphology. *Quaternary Science Reviews* 21, 1935-2017.

Barnekow, L., Possnert, G., and Sandgren, P. 1998: AMS C-14 chronologies of Holocene lake sediments in the Abisko area, northern Sweden - a comparison between dated bulk sediment and macrofossil samples. Geologiska föreningens i Stockholm förhandlingar 120, 59-67.

Birkeland, B. J. 1951. Altere meteorologische beobachtungen in Ullensvang, Luftdruck und temperatur seit 100 Jahren. In Geofysiske publikasjoner. Det norske videnskaps-akademi i Oslo.

Blake, G. R., and Hartge, K. H. (1986). Bulk Density. *In* Methods of Soil Analysis. pp. 363-375. American Society of Agronomy-Soil Science Society of America.

Blott, S. J., and Pye, K. 2001: Gradistat: A grain size distribution and statistics package for the analysis of unconsolidated sediments. *Earth Surface Processes and Landforms* 26, 1237-1248.

Bond, G., Kromer, B., Beer, J., Muscheler, R., Evans, M. N., Showers, W., Hoffmann, S., Lotti-Bond, R., Hajdas, I., and Bonani, G. 2001: Persistent solar influence on North Atlantic climate during the Holocene. *Science* 294, 2130-2136.

Bradley, R. S., Hughes, M. K., and Diaz, H. F. 2003: Climate in Medieval time. *Science* 302, 404-405.

Broecker, W. S. 1991: The great ocean conveyor. *Oceanography* 4, 79-89.

Calvo, E., Grimalt, J. and Jansen, E. 2002: High resolution $U_{k/37}$ sea surface temperature reconstruction in the Norwegian Sea during the Holocene. *Quaternary Science Reviews* 21, 1385-1394.

Cronin, T. M., Dwyer, G. S., Kamiya, T., Schwede, S., and Willard, D. A. 2003: Medieval Warm Period, Little Ice Age and 20th century temperature variability from Chesapeake Bay. *Global and Planetary Change* 36, 17-29.

Crowley, T. J., and Lowery, T. S. 2000: How warm was the medieval warm period? *Ambio* 29, 51-54.

Dahl, S. O., Bakke, J., Lie, O., 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.

Dahl, S. O., and Nesje, A. 1994: Holocene glacier fluctuations at Hardangerjøkulen, central southern Norway: a high-resolution composite chronology from lacustrine and terrestrial deposits. *The Holocene* 4, 269-277.

Dahl, S. O., and Nesje, A. 1996: A new approach to calculating Holocene winter precipitation by combining glacier equilibrium-line altitudes and pine-tree limits: A case study from Hardangerjokulen, central southern Norway. *Holocene* 6, 381-398.

Dahl, S. O., Nesje, A., Lie, O., Fjordheim, K., and Matthews, J. A. 2002: Timing, equilibrium-line altitudes and climatic implications of two early-Holocene glacier readvances during the Erdalen Event at Jostedalsbreen, western Norway. *Holocene* 12, 17-25.

Dean, W. E. 1974: Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition: comparison with other methods. *Journal of Sedimentary Petrology* 44, 242-258.

DNMI. 1993a. Nedbørsnormaler 1961-1990. Det Norske Meteologiske Institutt, Klimaavdelingen.

DNMI. 1993b. Temperaturnormaler 1961-1990. Det Norske Meteologiske Institutt, Klimaavdelingen.

Elvehøy, H. 1998. Samanlikning av massebalanse på Hardangerjøkulen og Folgefonna. In Oppdragsrapport (Elvehøy, H., Ed.), pp. 1-27. Norges vassdrag- og energiverk.

Feddersen, H. 2003: Predictability of seasonal precipitation in the Nordic region. *Tellus* 55A, 385-400.

Griffey, N. J., and Worsley, P. 1978: The pattern of Neoglacial variations in the Okstindan region of northern Norway during the last three millennia. *Boreas* 7, 1-17.

Haldorsen, S. 1981: Grain-size distribution of subglacial till and its relation to glacial crushing and abrasion. *Boreas* 10, 91-105.

Haldorsen, S. 1983: Mineralogy and geochemistry of basal till and their relationship to till-forming processes. *Norsk Geologisk Tidsskrift* 63, 15-25.

Heiri, O., Lotter, A. F., and Lemcke, G. 2001: Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results. *Journal of Paleolimnology* 25, 101-110.

Hopkins, T. S. 1991: The GIN-Sea - A synthesis of its physical oceanography and literature review 1972-1985. *Earth-Science Reviews* 30, 175-318.

Hurrell, J. W. 1995: Decadal Trends in the North Atlantic Oscillation: Regional temperatures and precipitation. *Science* 269, 676-679.

Hurrell, J. W. (2003). *The North Atlantic oscillation : climatic significance and environmental impact*. American Geophysical Union, Washington, D.C.

Hurrell, J. W., Hoerling, M. P., Phillips, A., and Xu, T. 2003: Twentieth Century North Atlantic Climate Change Part I: Assessing Determinism. *Climate Dynamics*.

Haakensen, N. 1989: Akkumulasjon på breene i Norge vinteren 1988-89. *Været* 13, 91-94.

Jørgensen, P. 1977: Some properties of Norwegian tills. *Boreas* 6, 149-157.

Karlén, W. 1976: Lakustrine sediments and tree-line variations as indikators of climatic fluctuations in Lappland, northern Sweden. *Geografiska Annaler* 58 A, 1-34.

Karlén, W. 1981: Lacustrine sediments studies. A technique to obtain a continous record of Holocene glacier variation. *Geografiska Annaler* 63A, 273-281.

Kolltveit, O. (1953). *Jondal i gamal og ny tid.* Jondal Historielag, Jondal.

Leonard, E. M. 1985: Glaciological and climatic controls on lake sedimentation, Canadian Rocky Mountains. *Zeitscrift für Gletcherkunde und Glazialgeologie* 21, 35-42.

Lie, O., Dahl, S. O., and Nesje, A. 2003: A theoretical approach to glacier equilibrium-line altitudes using meteorological data and glacier mass-balance records from southern Norway. *Holocene* 13, 365-372.

Lie, O., Dahl, S. O., Nesje, A., Matthews, J. A., and Sandvold, S. In press: Holocene fluctuations of a polythermal glacier in high-alpine eastern Jotunheimen, central-southern Norway: a multi-site, multi-parameter approach on lacustrine sediments. *Journal of Quaternary Science Reviews*, In press.

Lowe, J. J., and Walker, M. J. C. 2000: Radiocarbon dating the last glacial-interglacial transition (Ca. 14-9 C-14 ka BP) in terrestrial and marine records: The need for new quality assurance protocols. *Radiocarbon* 42, 53-68.

Matthews, J. A., Dahl, S. O., Nesje, A., Berrisford, M. S., and Andersson, C. 2000: Holocene glacier variations in central Jotunheimen, southern Norway based on distal glaciolacustrine sediment cores. *Quaternary Science Reviews* 19, 1625-1647.

Matthews, J. A., and Karlén, W. 1992: Asynchronous neoglaciation and Holocene climatic change reconstructed from Norwegian glaciolacustrine sedimentary sequences. *Geology* 20, 991-994.

Menounos, B. 1997: The water content of lake sediments and its relationship to other physical parameters: an alpine case study. *The Holocene* 7, 207-212.

Nesje, A. 1992: A Piston Corer for Lacustrine and Marine-Sediments. *Arctic and Alpine Research* 24, 257-259.

Nesje, A., and Dahl, S. O. 2001: The Greenland 8200 cal. yr BP event detected in loss-on-ignition profiles in Norwegian lacustrine sediment sequences. *Journal of Quaternary Science* 16, 155-166.

Nesje, A., Dahl, S. O., Andersson, C., and Matthews, J. A. 2000a: The lacustrine sedimentary sequence in Sygneskardvatnet, western Norway: a continuous, high-resolution record of the Jostedalsbreen ice cap during the Holocene. *Quaternary Science Reviews* 19, 1047-1065.

Nesje, A., Dahl, S. O., and Løvlie, R. 1995: Late Holocene glacier and avalanche activity in the Ålfotbreen area, western Norway: evidence from a lacustrine sedimentary record. *Norsk Geologisk Tidsskrift* 75, 120-126.

Nesje, A., Kvamme, M., Rye, N., and Løvlie, R. 1991: Holocene glacial and climate history of the Jostedalsbreen region, western Norway; evidence from lake sediments and terrestrial deposits. *Quaternary Science Reviews* 10.

Nesje, A., Lie, O., and Dahl, S. O. 2000b: Is the North Atlantic Oscillation reflected in Scandinavian glacier mass balance records? *Journal of Quaternary Science* 15, 587-601.

Meierding, T.C. 1982. Late Pleistocene glacier equilibrium-line in the Colorado Front Range: a comparison of methods. *Quaternary Research* 18, 289-310.

Nesje, A., Matthews, J. A., Dahl, S. O., Berrisford, M. S., and Andersson, C. 2001: Holocene glacier fluctuations of Flatebreen and winter-precipitation changes in the Jostedalsbreen region, western Norway: evidence from pro-glacial lacustrine sediment records. *Holocene* 11, 267-280.

Porter, S. C. 1975: Equilibrium-line altitudes of late quaternary glaciers in the Southern Alps, New Zealand. *Quaternary Research* 5, 27-47.

Roland, E., and Haakensen, N. 1985. 'Glasiologiske undersøkelser i Norge 1982', pp. 1-85. Norges vassdrags- og elektrisitetsvesen, Vassdragsdirektoratet, Hydrologiskavdeling.

Rosqvist, G. C. 1995: Proglacial lacustrine sediments from El Altar, Ecuador: evidence for late-Holocene climate change. *The Holocene* 5, 111-117.

Shabbar, A., Huang, J. P., and Higuchi, K. 2001: The relationship between the wintertime North Atlantic Oscillation and blocking episodes in the North Atlantic. *International journal of climatology* 21, 355-369.

Sigmond, E. M. O. 1985. Bedrock map of Norway. Norwegian geological survey, Trondheim.

Snowball, I. F., and Sandgren, P. 1996: Lake sediment studies of Holocene glacial activity in the Kårsa valley, northern Sweden: contrasts in interpretation. *The Holocene* 6, 367-72.

Souch, C. 1994: A methodology to interpret downvalley lake sediments as records of Neoglacial activity: Coast Mountains, British Colombia, Canada. *Geografiska Annaler* 76A, 169-185.

Stuvier, M., Reimer, P. D., Bard, P. J., Beck, J. W., Burr, G. S., Hughen, K. A., Kromer, B., McCormac, G., VanderPlicht, J., and Sprunk, M. 1998: INTCAL98 radiocarbon age calibration, 24,000-0 cal BP. *Radiocarbon* 40, 1041-83.

Sundborg, A. 1956: The River Klarälven, a study of fluvial processes. *Geografiska Annaler* 38, 125-316.

Sutherland, D. 1984: Modern glacier characteristics as a basis for inferring former climates with particular reference to the Loch Lomond stadial. *Quaternary Science Reviews* 3, 291-309.

Tvede, A. 1972. "En glasio-klimatisk undersøkelse av Folgefonni." Unpublished Thesis, Department of Geography, University of Oslo.

Tvede, A. M. 1979. Likninger til beregning av nettobalansen fra værdata. In Glasiologiske undersøkelser i Norge 1978 (Wold, B., and Repp, K., Eds.), pp. 71. Norges Vassdrags og elektrisitetsvesen.

Vorren, T. O. 1977: Grain-size distribution and grain-size parameters of different till types on Hardangervidda, south Norway. *Boreas* 6, 219-227.

Webb, P. W., and Orr, C. (1997). *Analytical Methods in Fine Particle Technology*. Micromeritics Instrument Corporation, Norcross, GS, USA.

Østrem, G. (1975). Sediment transport in glacial meltwater streams. *In* Glaciofluvial and glaciolacustrine sedimentation. (Jopling , A., and Mc Donald, B., Eds.), pp. 101-122. Society of Economic Palaeontologists and mineralogists. Special Publication.

Østrem, G., Dale Selvig, K., and Tandberg, K. 1988. Atlas of glaciers in South Norway. Norges vassdrags- og energiverk, vassdragsdirektoratet.