Utilizing physical sediment variability in glacier-fed lakes for continuous glacier reconstructions during the Holocene, northern Folgefonna, western Norway

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Abstract: The maritime plateau glacier of northern Folgefonna in western Norway has a short (subdecadal) response time to climatic shifts, and is therefore well suited for reconstructing high-resolution glacier fluctuations. The reconstruction presented here is based on physical parameters of glaciolacustrine sediments retrieved from two glacier-fed lakes and a peat bog north of the ice cap. Bulk density and modelled glacier net mass balance for the last 200 years show a remarkably similar pattern, where maximum sediment yield lags the glacier net mass balance by ~10 years. 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 maritime western Norway are therefore considered indicative of the strength of the westerly airflow associated with NAO during the Holocene. The early phase of mid-Holocene glacier growth (5200 cal. yr BP) was characterized by gradual glacier expansion culminating in the first Subatlantic glacial event at 2300 cal. yr BP. The climate during the last 2200 years has favoured increased glacier activity at Folgefonna. High-amplitude shifts in ELA may be explained by unstable modes of the westerlies causing significant variability of winter precipitation. During the last 2000 years, Folgefonna expanded and decayed with significant decadal variability. During the latest period of the ‘Mediaeval Warm Epoch’, Folgefonna advanced. The Neoglacial maximum, however, was reached during the ‘Little Ice Age’ at AD 1750 and AD 1870. The northern Folgefonna glacial record is compared to other Holocene glacier records in Scandinavia.

Key words: Glacier fluctuations, North Atlantic Oscillation (NAO), lake sediments, bulk density, ELA reconstructions, Folgefonna, glacier mass balance, Holocene, 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 climatic variability (e.g., Sutherland, 1984; Dyurgerov, 2002). Because studies on modern Norwegian glaciers have shown that sediment yield in glacier-fed lakes is positively correlated with glacier size (Roland and Haakensen, 1985) the proportion of glaciogenic material 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; Rosqvist et al., 2004). Various approaches use a conceptual model of glacier-meltwater induced sedimentation in which the minerogenic (nonorganic) component of the
sediments automatically 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., 2005). The most common approach is to use the organic content (loss-on-ignition (LOI) and total organic carbon (TOC)) as an inverse indicator of inorganic deposition. In lakes with high minerogenic sedimentation and/or low organic production (<5%) this approach has its limitations due to low ‘signal-to-noise’ ratio.

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 the position of the atmospheric polar front. The heat transport of the oceans on the west coast of Norway causes large temperature anomalies (Broecker, 1991; Hopkins, 1991). Large temperature gradients across the polar front 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; Hurrell et al., 2003; Nesje et al., 2000b). Atmospheric general circulation models have shown that the NAO is probably related to long-term trends in sea-surface temperatures (SST) (Feddersen, 2003; Hurrell et al., 2003). It is also demonstrated that higher winter precipitation in western Norway is related to stronger westerlies in the North Atlantic (associated with positive NAO weather modes) (Nordli et al., 2003). Mass balance, and hence size variations, of maritime glaciers in western Norway may thus be indicative for long-term trends of the westerlies.

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 to: (1) refine approaches for reconstruction of ELA variations using lake sediments with low organic/high minerogenic content; (2) evaluate strengths and weaknesses of sediment parameters used to obtain high-resolution ELA reconstructions; (3) reconstruct the Holocene glacial history of the plateau glacier Folgefonna at high temporal resolution; and (4) 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 of northern Folgefonna (23 km$^2$) is the seventh largest glacier in Norway. With its circular configuration it
ranges from 1644 to 1200 m and has a modern mean ELA of \(1465 \text{ m} \) (accumulation-area-ratio (AAR) \(0.7\)). The ice cap has five major outlet glaciers, Jordalsbreen, Jukladalsbreen (Figure 1), Botnabreen, Dettebrea and Juklavassbreen. About 12 km\(^2\) of the northern Folgefonna glacier drain northward (Figure 1), Botnabreen, Dettebrea and Juklavassbreen. About 12 km\(^2\) of the northern Folgefonna glacier drain northward (Figure 1), Botnabreen, Dettebrea and Juklavassbreen.

The ice cap occupies altitudes between 1200 and 1644 m with an average gradient of 100 m/km. The climatic response time of the glacier is 10–12 years, based on the positive mass balance years from AD 1989–92 which resulted in a glacial advance in AD 2001. In comparison, Jukladalsbreen has an altitudinal range between 1250 and 1644 m resulting in an average gradient of 150 m/km. The response time of this glacier is unknown, but is probably shorter than Jordalsbreen considering the steeper surface gradient. Based on the gradient of the glaciers, and the ice-flow velocity (\(~40 \text{ m/yr}^{-1}\)) (authors’ unpublished data), it is assumed that the sediment storage time in the glacier is short (\(~1–3 \text{ yr}^{-1}\)). The plateau glacier of northern Folgefonna has no supraglacial and limited englacial transport of glacial sediments.

The bedrock in the upper Jondal catchment consists mainly of acid meta-andesite, meta dacite, quartzite, migmatic and migmatic schist of Precambrian age (Sigmund, 1985; Askvik, 1995). The combination of acid rocks and the treeline situated at 600 m a.s.l. 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. The absence of superficial deposits reduces the influence of a paraglacial contribution to the glacial-fed lakes (Ballantyne and Benn, 1994; Ballantyne, 2002).

Using two meteorological stations along Hardangerfjorden (Station no. 4949, Ullensvang Forsøkså, 12 m a.s.l., 1962–88; Station no. 5013, Omastrand, 1 m a.s.l., 1962–90) (DNMI, 1993b), the present mean summer temperature (\(T_s\)) from 1 May to 30 September is suggested to be 12.7°C at sea level in Jondal. With an environmental lapse rate of 0.6°C/100 m (e.g., Sutherland, 1984) the mean \(T_S\) is close to 4.0°C at the modern ELA (1465 m) of northern Folgefonna. A local meteorological station 11 km from the present glacier terminus (Station no. 5696, Kvale, 342 m a.s.l., 1961–90) (DNMI, 1993a) records the 1961–90 mean winter (1 October to 30 April) precipitation (\(P_W\)) to have been 1434 mm. Based on an empirical, exponential increase in winter precipitation of 8%/100 m (Haakensen, 1989), the corresponding \(P_w\) at the ELA of northern Folgefonna is c. 3350 mm.

A peat bog, Hestadalsmyra, with bedrock thresholds covers an area of 0.05 km\(^2\) and is situated between the river from northern Folgefonna in the east and a small river from Hestadalsbotnen in 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. The mire was cored with a 110 mm PVC tube that was hammered into the mire and then excavated. The core was brought to the laboratory for radiocarbon dating and for magnetic susceptibility (MS) measurements.

Dravladalsvatn (938 m) covers an area of 1.35 km\(^2\) (Figures 1 and 2) [bathymetry from Statkraft, unpublished]. The lake is situated in a glacially eroded bedrock basin with the longest axis (2.5 km) oriented north/south. This particular glacial-fed lake is bound to receive meltwater wherever the glacier on the northern Folgefonna plateau is present. The distal, eastern basin, which was cored for this study, only receives the finest fractions (all sediments passed through a 125 \text{ mm} sieve) of these sediments, as the basin is sheltered from the main river current. 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 hydroelectric power.

Vassdalsvatn (490 m) covers an area of 0.17 km\(^2\) (Figure 1) [bathymetry presented in Bakke et al., 2005], and is the seventh glacial-fed lake downstream from northern Folgefonna. The lake is located in a glacially eroded bedrock basin and receives input of glacier meltwater-induced sediments at the present. The site is suitable to record major flooding events in the catchment and to register whenever Folgefonna has been present.

**Research approach and methods**

The reconstruction of Holocene glacier fluctuations at northern Folgefonna is based upon the following methods.

(1) Glacial-geomorphological mapping of the upper Jondal catchment, with special emphasis on former marginal moraines, glacier-meltwater channels and various ice-flow indicators.

(2) Dating of the ‘Little Ice Age’ (AD 1650–1930) glacial maximum by the use of lichenometry for absolute age estimate (data presented in Bakke et al., 2005). Lichenometry was also used to sort out the moraines older than ‘Little Ice Age’ (lichen diameter >200 mm).

(3) Calculations of former equilibrium-line altitude (ELA) are obtained using an AAR of 0.7 (Porter, 1975) on the advanced glaciers reconstructed by marginal moraines. 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.

(4) Ensuring a continuous record of former glacier size, two lacustrine sediment records retrieved from downstream distal glacial-fed lakes, were complemented with a peat bog stratigraphy.

The two glacial-fed lakes were cored using a modified piston corer taking up to 6 m long cores with diameter 110 mm (Nesje, 1992). In Dravladalsvatn, an ITH gravity corer was used to retrieve sediments from the uppermost part of the lake sediments. The laboratory analyses of the cores from Dravladalsvatn included (sampling resolution on all parameters = 0.5 cm) magnetic susceptibility (MS), weight loss-on-ignition (LOI) (Dean, 1974; Heiri et al., 2001), dry bulk density (g/cm\(^3\)) (DBD), water content and grain-size analysis using a Micromeretics Sedigraph 5100 (X-ray determination) (MasterTech, 1993). Grain-size statistics were performed by Gradistat 4.0 (Blott and Pye, 2001). Analyses of the two cores from Vassdalsvatn include LOI and MS.

Twenty-five bulk samples and four plant macrofossil samples were AMS dated from Dravladalsvatn and Vassdalsvatn. Terrestrial plant macrofossils for AMS radiocarbon dating were very sparse or absent in the lakes. A hard-water reservoir effect is not considered to produce erroneous dates since both lakes are located within acid Precambrian granite gneiss (see Barnekow et al., 1998; Lowe and Walker, 2000). The radiocarbon dates are shown in Table 1, and calibrated (cal. yr BP) according to INTCAL 98 (Stuiver et al., 1998). The intercepts used are based on the mean intercept if there is more than one calculated in CALIB 4.4.
Results

Moraine chronology

Marginal moraines in front of the outlet glaciers from northern Folgefonna indicate up to eight successively smaller glacier halts or advances/advances (Figure 1). The moraine chronology is not synchronous around northern Folgefonna. This may be due to differences in aspect and slope at Jordalssbreen and Jukladalsbreen.

The terminal moraines are marked with site names 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 (with lichen sizes over 200 mm) suggest a depositional age during the Lateglacial or early Holocene (Bakke et al., 2005). Terminal moraines (Ju-P1, Ju-P2, Ju-N1, Ju-N2 and Ju-N3) north of Lake Jukladalsvatn demonstrate that Jukladalsbreen has crossed the valley several times during the Holocene. Historical sources indicate river-flooding events in Gressdal (upper Jondal catchment) during the ‘Little Ice Age’ that caused damage to the surrounding farmland (Kolltveit, 1953). Glacier advances of Jukladalsbreen led to the formation of a large lake to the east of the glacier that later was catastrophically drained when the glacier retreated and/or the water pressure became higher than the ice pressure. These floods can be classified as Jokulhlaups that are reported also from the southern part of the Folgefonna glacier (Tveide, 1989). 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 lake.

In the cirque Hestadalsbotnen (no glacier at present) there are four moraine ridges (Figure 1). H-P1 and H-P2 are assumed to be of Preboreal age (11 500–9950 cal. yr BP) (lichen = > 200 mm) and H-1 and H-2 were, according to the lichen measurements, deposited AD ~ 1870 and ~ 1750, respectively. The glacial activity in Hestadalsbotnen is, based on lichenometry, inferred to have been synchronous with the northern Folgefonna.

Lithostratigraphy and radiocarbon dates

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 14C 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 14C 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 14C yr BP (T-13600). The upper unit (unit A) consists of humus with grass on the top.

Dravladalsvatn

The interpreted lithostratigraphies from the individual cores are shown in Figure 4. 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 of grey silt and clay (unit G) below a gradually transition (unit F) into dark brown gyttja (unit E). The transition between unit G and F is dated at 8645 ± 70 14C 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 14C yr BP (Poz-3177) and of 5530 ± 40 14C 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 14C yr BP (TUa-3628A). The upper part of the unit was radiocarbon dated at 2000 ± 40 14C 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 14C 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. DDB 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 was missing.
Table 1  Radiocarbon dates obtained from the cores studied. When more than one calibrated intercept age is given, the mean intercept value is used.

<table>
<thead>
<tr>
<th>Site</th>
<th>Lab. no.</th>
<th>Depth (cm)</th>
<th>Type of material</th>
<th>Radiocarbon age ± 1 sigma (cal. yr BP)</th>
<th>Intercept ± 1 sigma (cal. yr BP)</th>
<th>± 2 sigma (cal. yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Vassdalsvatn Core I</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Vassdalsvatn I</td>
<td>Beta-102930</td>
<td>28–31</td>
<td>Gyttja</td>
<td>1150 ± 70</td>
<td>1060</td>
<td>1170–970</td>
</tr>
<tr>
<td>Vassdalsvatn I</td>
<td>Beta-102931</td>
<td>117–120</td>
<td>Gyttja</td>
<td>2280 ± 60</td>
<td>2240</td>
<td>2350–2160</td>
</tr>
<tr>
<td>Vassdalsvatn I</td>
<td>Beta-102932</td>
<td>182–185</td>
<td>Gyttja</td>
<td>3370 ± 70</td>
<td>3645</td>
<td>3690–3480</td>
</tr>
<tr>
<td>Vassdalsvatn I</td>
<td>Beta-102933</td>
<td>250–253</td>
<td>Gyttja</td>
<td>4270 ± 80</td>
<td>4810</td>
<td>4965–4650</td>
</tr>
<tr>
<td>Vassdalsvatn I</td>
<td>Beta-102934</td>
<td>295–298</td>
<td>Gyttja</td>
<td>5200 ± 70</td>
<td>5960</td>
<td>6170–5750</td>
</tr>
<tr>
<td>Vassdalsvatn I</td>
<td>Beta-102935</td>
<td>368–372</td>
<td>Gyttja</td>
<td>8260 ± 80</td>
<td>9245</td>
<td>9415–9130</td>
</tr>
<tr>
<td>Vassdalsvatn I</td>
<td>Beta-102936</td>
<td>525–535</td>
<td>Gyttja</td>
<td>8330 ± 50</td>
<td>4910</td>
<td>4965–4840</td>
</tr>
</tbody>
</table>

| **Vassdalsvatn Core II**    |          |            |                  |                                        |                                 |                        |
| Vassdalsvatn II             | T-13607  | 19         | Gyttja           | 2100 ± 85                              | 2050                            | 2295–1970              |
| Vassdalsvatn II             | T-13608  | 83–84      | Gyttja           | 1900 ± 70                              | 1840                            | 1920–1735              |
| Vassdalsvatn II             | T-13788A | 77–79      | Gyttja           | 2310 ± 60                              | 2080                            | 2360–2160              |
| Dravladalsvatnet Core I     |          |            |                  |                                        |                                 |                        |
| Dravladalsvatnet Core I     | Poz-3175 | 1          | Macro fossil     | 2060 ± 30                              | 2030                            | 2060–1990              |
| Dravladalsvatnet Core I     | TUa-3628A | 57         | Macro fossil     | 2315 ± 45                              | 2310                            | 2355–2305              |
| Dravladalsvatnet Core I     | Poz-3176 | 72         | Macro fossil     | 5530 ± 40                              | 6340                            | 6395–6290              |
| Dravladalsvatnet Core II    | Poz-3177 | 82         | Gyttja           | 8090 ± 40                              | 9055                            | 9220–9000              |
| Dravladalsvatnet Core II    | TUa-3629A | 88         | Gyttja           | 8645 ± 70                              | 9660                            | 9690–9540              |

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 14C yr BP (TUa-3632A) and in the upper part 5050 ± 70 14C yr BP (Poz-3256). Unit B is a transitional unit with a change from dark brown gyttja to grey silt gyttja. A radiocarbon date in the lower part yielded an age of 1910 ± 45 14C yr BP (TUa-3630). The upper part of unit A, three inverted radiocarbon dates were obtained, yielding 2315 ± 25 14C yr BP (Poz-3798), 2320 ± 45 14C yr BP (TUa-3640A) and 2565 ± 30 14C yr BP (Poz-3178), respectively, toward the top of the core.

A short 39 cm gravity core (ITH) from Lake Dravladalsvatn consisted of silt and clay with LOI values below 6% throughout the entire core. Grain-size analyses, DBD and MS showed high-frequent fluctuations. No radiocarbon dates have been obtained from this core (Figure 7).

**Vassdalsvatn**

In Vassdalsvatn, the two cores showed remarkably different lithostratigraphy. Core I (550 cm) in Vassdalsvatn was retrieved in the central part of the lake, 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 inverted radiocarbon dates were obtained, yielding 2315 ± 2155 14C yr BP (Poz-3179). Unit A consisted of grey clay and silt with an age of 3215 ± 60 14C yr BP (TUa-3631A). At 78 cm a radiocarbon date yielded an age of 1910 ± 45 14C yr BP (TUa-3630). In the upper part of unit A, three inverted radiocarbon dates were obtained, yielding 2315 ± 25 14C yr BP (Poz-3798), 2320 ± 45 14C yr BP (TUa-3640A) and 2565 ± 30 14C yr BP (Poz-3178), respectively, toward the top of the core.

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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 $6280 \pm 60 \text{ C yr BP}$ (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 $3820 \pm 50 \text{ C yr BP}$ (UtC-6694) below and $3460 \pm 60 \text{ C yr BP}$ (UtC-6693) above. From this unit several radiocarbon dates have been obtained, $3230 \pm 40 \text{ C yr BP}$ (UtC-6692), $2765 \pm 45 \text{ C yr BP}$ (UtC-6691), $1900 \pm 70 \text{ C yr BP}$ (T-13608) and $2310 \pm 60 \text{ C yr BP}$ (T-13788A), respectively (Figure 5B). The two layers with plant macrofossils and fine sand (unit D and B) were radiocarbon dated with one sample below each unit. Unit D yielded an age of $1900 \pm 70 \text{ C yr BP}$ (T-13608), whereas unit B yielded an age of $2100 \pm 85 \text{ C yr BP}$ (T-13607).

**Age-depth relationship**

The age-depth models for Dravladalsvatn and Vassdalsvatn are shown in Figure 6. Both models are constructed by linear interpolation between radiocarbon dates or between ‘fixed’ points in the cores. A major problem establishing the age-depth relationship 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 erosion and re-deposition of terrestrial plant material during river floods. The draining of the glacier-dammed lake in Jukladalen may have led to raised lake level in Dravladalsvatn and thereby erosion along the shores. Another possible explanation for the inverted AMS bulk dates in the upper part could be the low organic content in the samples. Despite the problematic radiocarbon dates, the age-depth model in Dravladalsvatn is constrained by radiocarbon dates in Vassdalsvatn and Hestadalsmyra by correlation to flood events inferred from analyses of sorting and mean used as time markers (Arnaud *et al.*, 2002). Periods with poorer sorting were interpreted as events with abrupt change in the input of minerogenic sediments in Dravladalsvatn (Figure 8), interpreted as river-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 (‘fixed points’) in the age-depth models. In Vassdalsvatn, the flooding events were represented by layers of sand and inwash of plant macrofossils. The sites belong to the same drainage system, and these large floods should therefore be detectable in both lakes. In the peat bog Hestadalsmyra silt layers were interpreted to reflect periods with enhanced glacial activity in the cirque Hestadalsbotnen. It is assumed that glacial activity in Hestadalsbotnen was synchronous with periods of glacier growth at northern Folgefonna. The major flooding events occurred at $\sim$3500 cal. yr BP (age from core I in Vassdalsvatn and core II in Dravladalsvatn), $\sim$2250 cal. yr BP (age from core II in Dravladalsvatn, core II from Vassdalsvatn and core II in Dravladalsvatn), $\sim$1050 cal. yr BP (ages from mire Hestadalsmyra and core II in Dravladalsvatn). 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 mean sorting, MS, DBD and LOI. The advantage of the gravity core is that the core contains an undisturbed section of the top sediments in Dravladalsvatn. The uppermost mean—sorting anomaly is correlated to a historical documented flood in the catchment (7 cm in core = AD 1890) (Kolltveit, 1953). This gives the gravity core the same sedimentation rate as the upper part of core II ($\sim$10 yr/cm$^1$), the remaining 32 cm of the core is assumed to have the same sedimentation rate (Figure 7). The coring sites for these two cores are at 56 m water depths, only a few metres apart.

The age-depth model for the Vassdalsvatn record is complicated by two inverted radiocarbon dates in the upper part of core II probably due to inwash of terrestrial plant macrofossils during river-flooding events. The basal date in core I is apparently too young, probably due to contamination inflow when the core was jacked out of the sediments.

**Discussion**

The minerogenic input to the eastern basin of Dravladalsvatn is suggested to represent a ‘pure’ glacial signal as there is only a sparse cover of colluvium and basal till around the lake. The small bouldery colluvial fans surrounding Dravladalsvatn are inferred to deliver different grain sizes than those produced by the glacier due to the short transport length and their possible influence is considered to have been minimal. Subaquatic erosion of previously deposited glacigenic sediments is not likely due to the low energy in this eastern part of the lake. The bedrock threshold and the position of inlet and outlet in the western main basin make the coring sites favourable for recording periods when there were minerogenic sediments in suspension.
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 DBD/MS records.

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 years before present. As seen from the depth scale, there was a notable change in sedimentation rate around 130 cm. Higher ‘sorting’ values mean poorer sorting of the sediments. Grey shaded areas show sorting anomalies, and dark grey shaded areas show anomalies used for constructing the age-depth models.
The composite chronostratigraphy from Dravladalsvatn is based on the age-depth model in Figure 7. Due to the large water depth (~70 m) (artificial rise of lake level due to hydroelectric power plant construction), it was difficult to retrieve the uppermost soft sediments. However, reliable correlations between the core stratigraphy 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 difference in sedimentation rates between the two piston cores is caused by difference in water depth and distance to the bedrock threshold. Core I was taken at 35 m water depth in the distal part of the basin whereas core II was taken at 55 m water depth (natural water level) closer to the bedrock threshold. Despite the inverted radiocarbon dates in Dravladalsvatn, the age-depth model of the composite stratigraphy is constrained by radiocarbon dates from Vassdalsvatn and Hestadalsmyra together with a historically documented flood event.

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

The grain-size distribution in Dravladalsvatn is shown in Figure 9. Generally, there was a high input of coarse particles during timespans when the sediments were dominated by gyttja. The very coarse silt fraction is interpreted to be sediments from the surrounding catchment and in this context regarded as ‘noise’. As the LOI values decrease, suggested glacigenic sediments consisting of clay and fine silt dominates 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. The positive correlation probably reflects an increased energy in the lake. As the glacier grew the water discharge increased, transporting more sediment over the bedrock threshold and into the coring site. The negative correlations indicate that the coarser grain sizes in Dravladalsvatn are transported into the lake by slopewash from the lake surroundings. Based on these results, the lithostratigraphy was divided into four phases (Figures 7 and 9).

MS shows low values when gyttja dominated the sediments (phases III and II), whereas the MS values increased rapidly as the proportion of minerogenic sediments increased in phase IV. During phase I the MS signal rose as the DBD values increased, reflecting varying influx of minerogenic 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. Grain-size and DBD analyses are not performed for the cores from Vassdalsvatn. The differences in lithostratigraphy between the two cores may be explained by the coring sites lying with different distance from the inlet and outlet of the main river. The lithostratigraphy in Vassdalsvatn was used to complement Dravladalsvatn regarding major flooding events, represented by sandy organic-rich layers.

**Bulk density as a proxy for glacier size**

Loss-on-ignition has traditionally been used as an inverse indicator for inorganic lake sedimentation. The approach is widely used to reconstruct glacier variations (e.g., Karlén, 1976; 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 very low. This approach has therefore natural limitations in high alpine and
Arctic lakes. In Dravladalsvatn, the LOI values were below 5% during periods when northern Folgefonna was at its present size (Figure 10). It was 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).

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 glaciolacustrine sediments (e.g., Østrem, 1975). The use of grain-size variations have, however, not been widely used in this context. An important factor concerning the grain-size distribution in glacial-fed 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 glacial-fed lakes are therefore mainly reflecting changes in fluvial and lacustrine systems. High-energy streams deposit coarser sediments, and vice versa (e.g., Hjulström’s diagram; Sundborg, 1956). 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 commonly consists of grain sizes suitable for suspension (1–63 μm), commonly giving more sediments per time than an ‘open-ended’ lake basin.

By definition, bulk density expresses the ratio of the mass of dry solids to the bulk volume of the 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 glacial-fed 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). 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 cores I and II from Dravladalsvatn this relationship is very strong ($r^2 = 0.97$).

**ELA variations at northern Folgefonna**

A relationship between grain-size variations (sorting–mean anomalies), DBD and glacier size based on the analysis has been established (Figure 12). The altitudinal position of 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 (Figure 12B). The moraines Jo-1, Jo-2 and Jo-3 are independently ‘dated’ by the use of lichenometry and historical sources, whereas Ju-N1 and Ju-N2 are relatively dated by Schmidt hammer (Bakke et al., 2005). Periods with sorting anomalies (due to flooding events) have been removed from the ELA reconstruction (open squares in Figure 12).

![Figure 10](image-url)  
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 of inorganic sedimentation. Dotted line shows DBD. (B) Regression between residue (%) and DBD, showing a close relationship ($r^2 = 0.8$) between the two parameters.
The moraines in front of Jukladalsbreen indicate that the glacier had two advances (2250 cal. yr BP and 1750 cal. yr BP). Angular minerogenic particles are obtained from sediments dominated by gyttja and angular minerogenic particles.

The moraines in front of Jukladalsbreen indicate that the glacier had two advances (2250 cal. yr BP and 1750 cal. yr BP) which are dated by Schmidt hammer. The flood events detected in Vassdalsvatn before entering Dravladalsvatn (Figure 1). The uppermost gravity core of Jukladalsvatn indicate a rapid retreat of the glacier, as the glacier area gradually became bigger from AD 1800 to 1840 (Figure 13). The uppermost gravity core is undated; however, the age model is correct, another interesting feature is that there apparently 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.

The ELA reconstruction has been divided into six phases.

(1) Between 9660 and 5200 cal. yr BP the ELA at northern Folgefonna was above the highest mountain (>1550 m) 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 buildup 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 from 1440 m to 1360 m as the glacier area gradual became bigger.
(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 (model; n = 200, DBD; n = 40) at Folgefonna (Figure 13). The uppermost gravity core is undated; however, age control is established using linear interpolation between the sediment surface (present) and a historically dated flood (AD 1890) as a sorting–mean anomaly as a time marker. Figure 13 shows some equations for modelling of the glacier mass balance (Bw/Bn and Bn) of Folgefonna based on temperature and precipitation from the Bergen-Florida meteorological station. The equation was later reformulated (Elvebøy, 1998) and established also for the northern part of Folgefonna:

\[
B_n = 444 + 2.16 \times P - 54 \times T_1
\]

where \( P \) is winter precipitation in Bergen (01.10 – 31.05) and \( T_1 \) is average summer temperature (01.06 – 31.08). The equation gives high predictability compared to the net mass balance from 1963 to 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 to 1840 (Birkeland, 1932). As there is a lack of precipitation records for this timespan, 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 1840 reflect the retreat of the glacier after the AD 1750 glacier event, and may explain the high DBD values during the timespan. If the age model is correct, another interesting feature is that there apparently 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.

**Implications of climatic importance from the northern Folgefonna record**

The approach shown in this paper provides new methodological 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 14, 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 Bakke et al. (2005), the early-Holocene event chronology is discussed on the basis of lake sediments from Vetlavatn (Figure 1), 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 recognized 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).
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, glacial-fed and nonglacial 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.

(2) The first Neoglaciation at northern Folgefonna started around 5200 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 Neoglaciaation at northern Folgefonna has some notably consistent modes. The glacial advance from 5200 cal. yr BP until ~2300 cal. yr BP was a phase with gradual 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 Voring 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 indicate a change in the atmospheric and oceanic conditions, rather than abrupt climate changes as recorded during the early Holocene (Dahl et al., 2002; Nesje et al., 2004). The reduced SST in the North Atlantic and the expansion of the Folgefonna glacier may indicate that the ocean has a major forcing on the precipitation distribution in the North Atlantic Realm.

(5) The Holocene glacier record from northern Folgefonna indicates high-frequently 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 are also recognized at Jostedalsbreen (Nesje et al., 2001), Hardangerjøkulen (Dahl and Nesje, 1994) and at Bovtunbreen (Matthews et al., 2000) in southern Norway during the

![Figure 12](image-url)
same timespan. From northern Norway, two late-Holocene glacier readvances are recognized at Okstindan dated at 3000–2500 \(^{14}\text{C}\) yr BP and 1250–1000 \(^{14}\text{C}\) 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-frequency climatic shifts, leading to glacier expansion and decay, are representative for at least western Scandinavia. Folgefonna is a maritime glacier where ~80% of its modern net balance 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 (associated with positive NAO weather mode) that appears in a less stable mode than during the period from 5200 to 2200 cal. yr BP. Another explanation may be a 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 et al., 2003).

(6) The period termed ‘the Mediaeval 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 1300 (Cronin et al., 2003). It is no evidence for the temperature record exceeding 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, ~1870 and ~1930 with successively smaller glacier advances, but marked by distinct marginal moraines. This is in accordance with earlier studies from Folgefonna and also with historical sources 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 climatic 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%) due to low signal-to-noise ratio.
Sorting mean anomalies can be used to track abrupt changes in the sedimentation environment in a lake and thereby validate the use of lake sediments for reconstruction of former glacier fluctuations.

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

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.

At 2200 cal. yr BP there was a significant change in glacier size, from a small glacier to glacier size larger than at present. The record from the last 2200 years shows high-frequently glacial fluctuations at decadal and centennial timescales. It is indicated that the so-called ‘Mediaeval Warm Epoch’ was a humid phase at northern Folgefonna, as glacier growth and decay during this timespan 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.

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. A comparison between DBD and modelled glacier net mass balance shows a remarkably similar pattern.

Dry bulk density (DBD) has the potential to resolve even small changes in silt production caused by interference in mass balance over short periods (subcentennial).

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