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

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

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

Introduction

Along the coast of Troms, northern Norway (Fig. 1), the northward heat advection of air and water masses into the Norwegian Sea region produces some of the largest temperature anomalies in the world, with a mean January air temperature about 24°C warmer than the latitudinal mean in Tromsø (69° 40° N) (Encyclopaedia Britannica, 2003). The regional temperature and precipitation patterns are sensitive to variations of this heat advection. If it is reduced / turned off, the region is suggested to experience an ‘instantaneous’ shift from anomalously warm and open ocean climate conditions to a wide-spread cold climate regime with continuous sea-ice cover. Fluctuations in glacier size and equilibrium-line altitude (ELA) are primarily the result of variations in mean ablation-season temperature and / or winter precipitation as snow. Because small glaciers have a short time delay on climatic changes (e.g. Nye, 1960; Paterson, 1994), fluctuations in glacier magnitude and ELA is suggested to be a sensitive proxy for variations in the atmosphere-ocean interaction in the Norwegian Sea region.

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

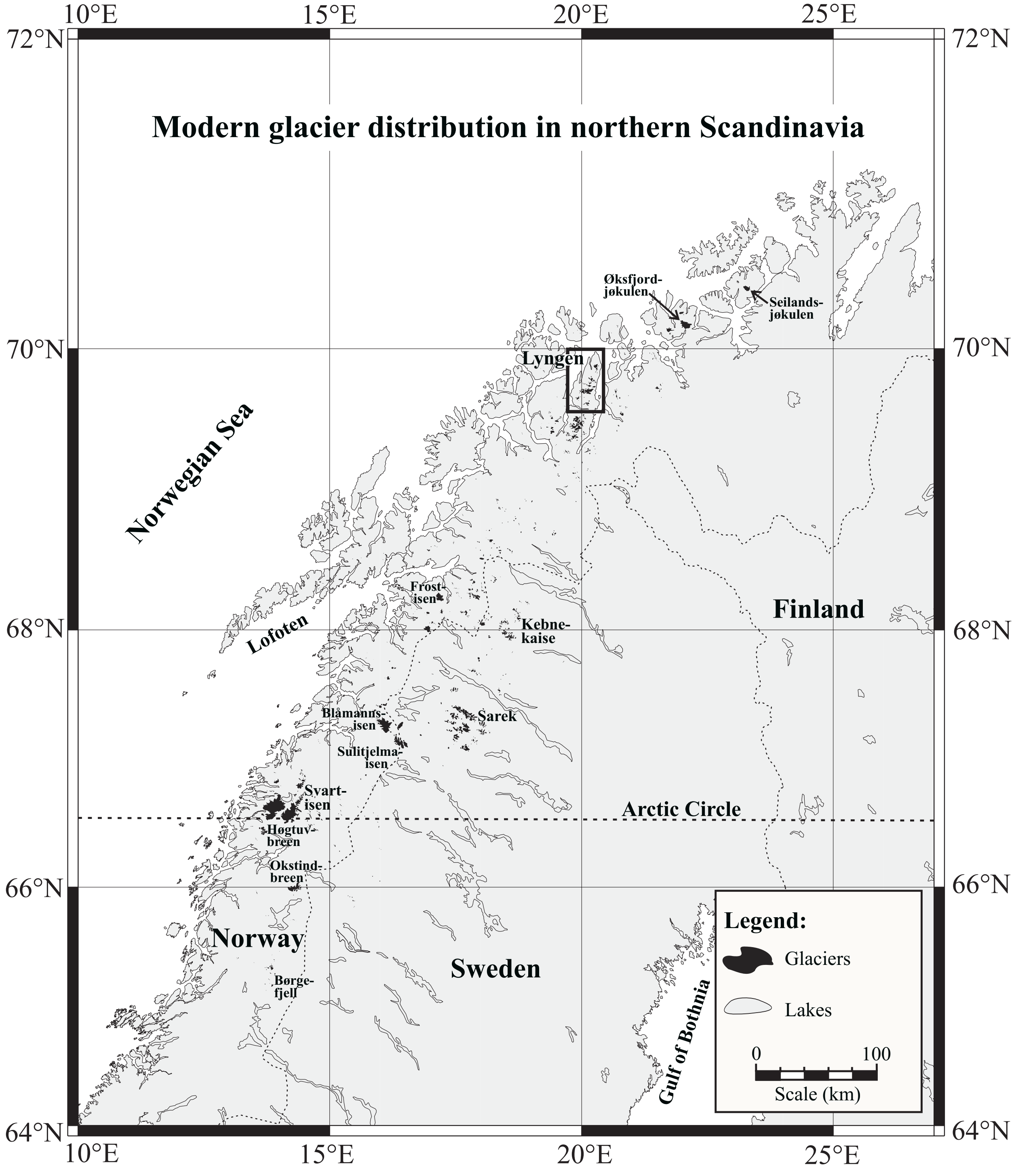

Figure 1 Map showing modern glacier distribution in northern Scandinavia. Bold frame shows the study area north of the fjord Kjosen on the Lyngen Peninsula. The continental glaciers around Kebnekaise and Sarek in northern Sweden have the highest equilibrium-line altitudes (ELAs), whereas glaciers at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

Fimreite et al., 2001; Plassen and Vorren, 2002; Vorren and Plassen, 2002), reconstruction of local glaciers (Karlén, 1973, 1976, 1981, 1988, 1997; Whalley, 1973; Karlén and Denton, 1975; Griffey, 1976; Worsely and Alexander, 1976; Corner, 1978; Griffey and Worsley, 1978; Gellatly et al., 1986, 1988, 1989; Gordon et al., 1987, 1996; Whalley et al., 1989, 1996; Ballantyne, 1990; Snowball, 1996; Snowball and Sandgren, 1996, 1997; Rea et al., 1998, 1999; Evans et al., 2002; Winkler et al., 2003), reconstruction of former sea-level changes (e.g. Marthinussen, 1960, 1962; Donner et al., 1977; Hald and Vorren, 1983; Møller, 1984, 1985, 1986, 1987, 1989; Corner and Haugane, 1993; Kverndal and Sollid, 1993), reconstruction of summer temperature by biological proxies (e.g. pollen, macrofossils, diatoms, chironomids) (e.g. Vorren et al., 1988; Alm, 1993; Vorren and Alm, 1999; Seppä and Birks, 2001, 2002; Korhola et al., 2002; Seppä et al., 2002a, 2002b), reconstruction of former tree-line altitude by megafossils (Kullman, 1995, 1999, and references therein), oxygen-isotope records (Berglund et al., 1996; Shemesh et al., 2001;

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

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

Study area

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

Except for ice caps at the summits of Jiehkkevàrri (altitude 1833 m) and Bàlggesvàrri (altitude 1625 m) south of the study area, cirque glaciers dominate the present glaciation on the Lyngen Peninsula. Large cirque glaciers like Sydbreen (c. 9 km^2) and Strupbreen (c. 8. 5 km^2), however, may be regarded as valley glaciers. High-altitude glaciers are probable polythermal (subpolar), as they are thin and exist within the zone of discontinuous permafrost, whereas valley-outlet glaciers from the ice caps and low-altitude cirque glaciers, reaching down to 400 m, are inferred to be temperate.

Lake Aspvatn (altitude 35 m) is the second proglacial lake c. 6 km downstream of two partly coalescing cirque glaciers named Lenangsbreene (c. 2 km^2) and three smaller cirque glaciers (c. 0.6 km^2) in western parts of Strupskardet (Figs. 2-4). Meltwater from these glaciers, including the dominant Lenangsbreene, drains through Blåvatn before it enters Aspvatn and then to the fjord Sør-Lenangen.

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

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

Methods

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

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

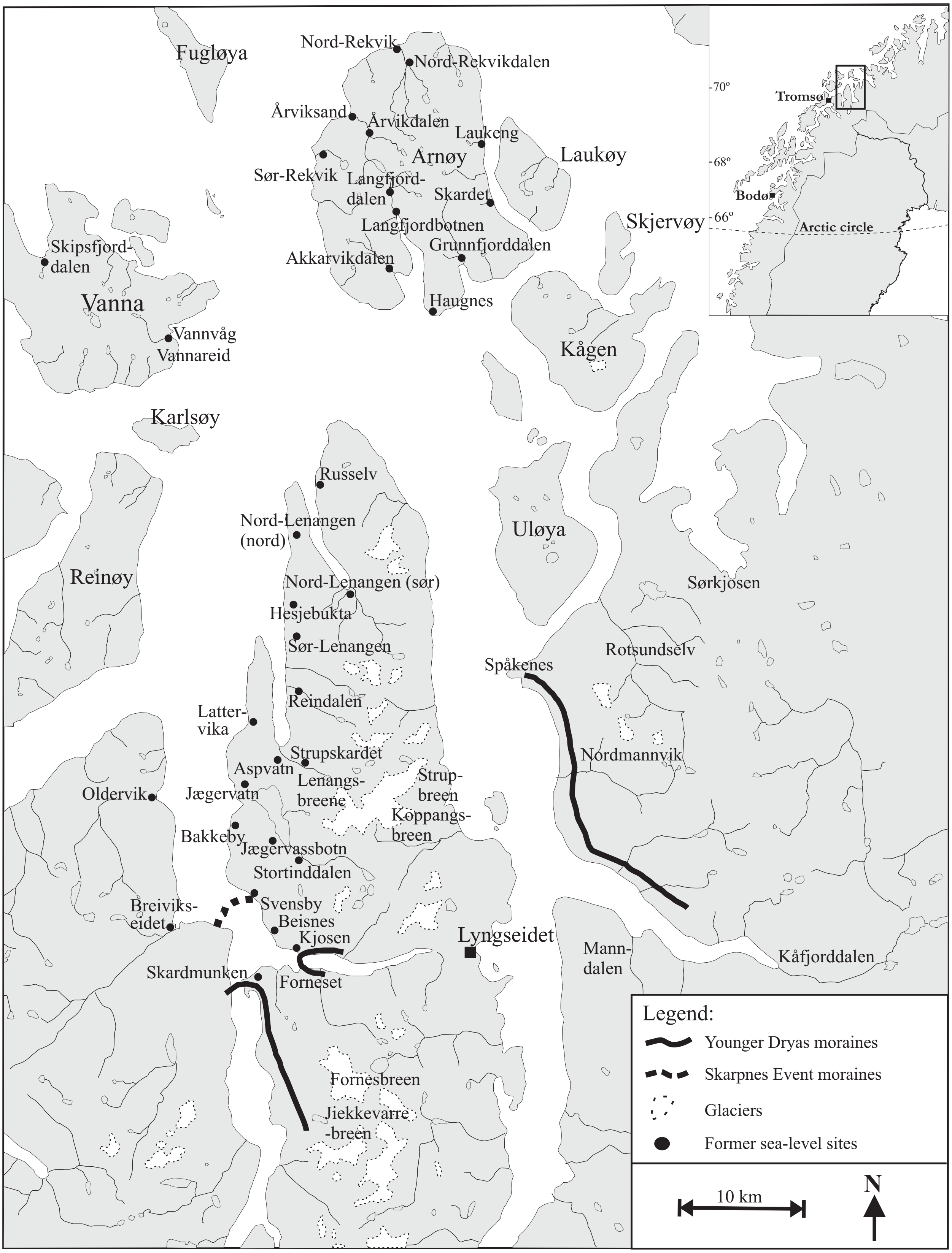

Figure 2 Map showing the northern part of the Lyngen Peninsula. Dotted lines show the suggested ice margins formed by the Scandinavian Ice Sheet during the Skarpnes event and solid lines indicate the ice margin(s) during the Tromsø-Lyngen event (Younger Dryas). Black dots indicate sites used in the equidistant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

sandurs), raised marine terraces, beach ridges and shorelines, rock glaciers and tundra polygons.
- The timing of Lateglacial events linked to sea level was sorted out based on the regional sea-level history recorded around Arnøy (Kverndal and Sollid, 1993) and from Lyngen (Corner and Haugane, 1993) and correlated with observations in this study from Russelv to Skardmunken (Fig. 2).

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

- Lichenometry based on *Rhizocarpon geographicum* (e.g. Ballantyne, 1990; Matthews, 1994) was used to establish relative-age chronologies for the marginal moraines in front of Lenangsbreene with particular emphasis of indentifying moraines older than the 'Little Ice Age' maximum in Lyngen.
- The deposition of distinct marginal moraines is related to (rather) short periods when the glacier is in steady-state, whereas younger glacier (e.g. the 'Little Ice Age') advances may have erased older marginal moraines (e.g. Matthews, 1991; Dahl and Nesje, 1994). Hence, proglacial sites (lacustrine and terrestrial) beyond the maximum extent of the suggested glacier advances within the studied time span are taken into account to obtain continuous records of glacier fluctuations. The methods related to proglacial sites are based on a conceptual model of glacier-meltwater induced sedimentation in which the minerogenic (nonorganic) component of the sediments is related to the presence of a glacier in the catchment (Karlén, 1981; Leonard, 1985; Dahl et al., 2003).
- Two cores were retrieved from Aspvatnet (Fig. 3) during the late summer of 2000 by use of a modified piston corer (Nesje, 1992) taking up to 6 m long tubes with diameter 11 cm. The two cores were 4.3 and 4.5 m long, respectively, and were taken from the deepest part of the lake. Both cores were subsampled at 0.5 cm intervals using volume determinant sampling equipment in the laboratory for a number of analyses:
- For loss-on-ignition (LOI) (Dean, 1974) the samples were ignited for 1 hour at 550ºC.
- Wet and dry bulk density (DBD) was measured before and after drying.
- Water content was calculated as per cent of dry sample weight.
- Fine-grained sediments were analysed using a 'Micromeretics SediGraph 5100 Particle Size Analyse System' (MasterTech, 1993). This system measures particle diameter based on Stoke´s law in a range from 300 to 0.1 micrometer (µm) equivalent spherical diameters by shooting x-rays through suspended material.
- Prior to sub-sampling, surface magnetic susceptibility (MS) was determined at 5 mm intervals with an in-house constructed automatic instrument using a Bartington MS2E sensor on a cleaned surface (2/3 of core) covered with a thin plastic film (<0.1 mm thick). A total of 97 cubic plastic boxes (2x2x1.8 cm) were retrieved. Sampling intervals varied from 2 cm (0-100 cm and 210-270 cm) and between 10 and 20 cm for the remaining core. Sample weights were converted to dry weight using the average water content value for parallel samples, whereas MS was determined on a KLY-2 induction balance (sensitivity

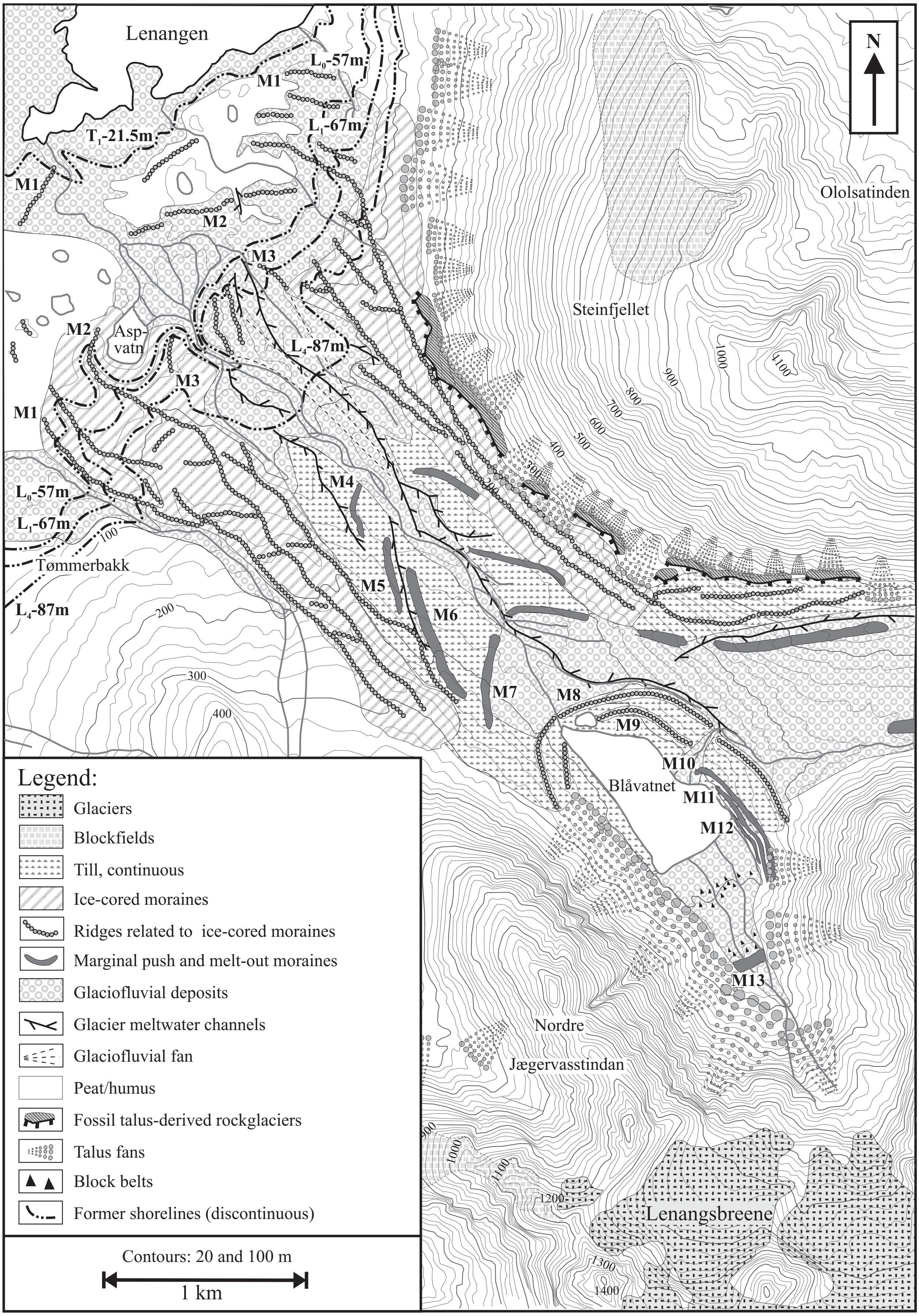

Figure 4 Glacial-geomorphological map showing the valley Strupskardet and the two proglacial lakes Blåvatnet and Aspvatnet. The valley floor is dominated by glaciofluvial deposits and till. The marginal moraines M1-M3 and M8-M9 are mapped as ice-cored moraines indicating continuous permafrost during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

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

- Glacier ELAs are reconstructed using an accumulation-area ratio (AAR) of 0.6 (Dahl et al., 1997). To reconstruct former ELAs of debris-mantled glaciers in high-mountain regions may be difficult (Clark et al., 1994; Benn and Lehmkuhl, 2000). Because of rather well-defined landforms, however, remnants of ice-cored marginal moraines were used to reconstruct former ELAs in the study area. The calculation of the glacier area distribution was carried out using the vector-based GIS program MapInfo 6.0 on an N-50 map datum.
- Winter precipitation estimates were calculated based a close exponential relationship between mean ablation-season temperature t (°C) (1 May–30 September) and mean solid winter precipitation A (m water equivalent) (1 October-30 April) at the ELA of Norwegian glaciers in maritime to continental climatic regimes (both temperate and polythermal glaciers) (Liestøl in Sissons, 1979; Ballantyne, 1989; Dahl et al., 1997, and references therein):

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

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

Results

The valley of Strupskardet and the neighbouring mountains contain a number of different active, inactive, and relict landforms that are related to climate change (Figs. 3 and 4). The mountains surrounding Strupskardet are rather flat and undulating with autochthonous blockfields on the summits and allochthonous blockfields at the steeper upper slopes. Along the northern valley face of Strupskardet and at both sides of Blåvatn there are active talus formation and several snow avalanche tracks. There is no active moraine formation in front of Lenangsbreen at present. Along the valley floor of Strupskardet, however, the main features are thirteen marginal moraines (M1-M13).

Moraine chronology

Moraines M1-M3

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

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

Moraines M4-M7

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

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

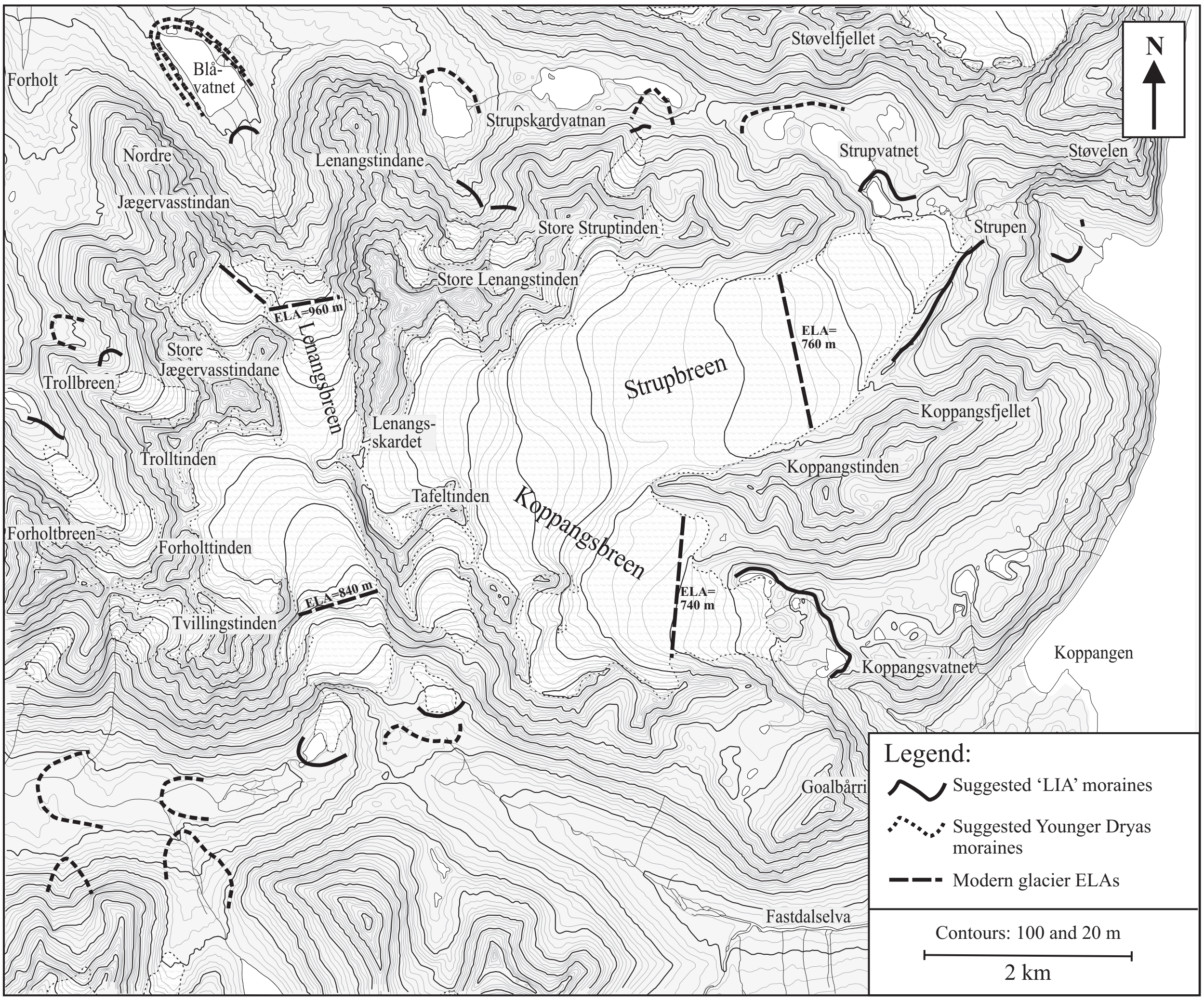

Figure 5 Map showing the modern TPW-ELAs and reconstructed ice-marginal moraines from the 'Little Ice Age' and Younger Dryas at the ice field consisting of Strupbreen, Lenangsbreene, Koppangsbreen and several smaller glaciers. The TPW-ELA is about 200 m lower at the eastern side of the Lyngen Peninsula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

Moraines M8-M9

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

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

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

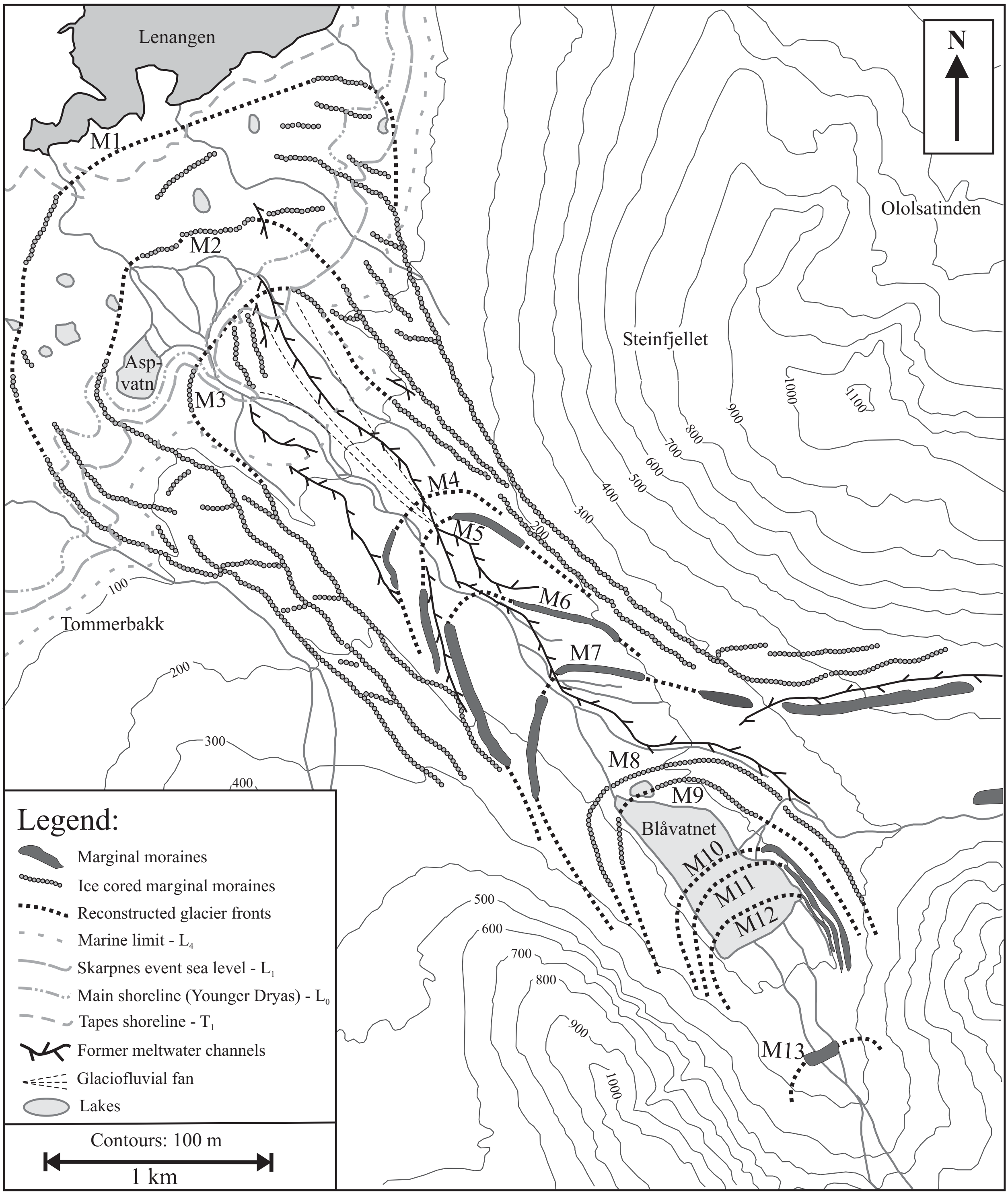

Figure 6 Map showing the reconstructed glaciers at thirteen glacier stages obtained from the mapped moraine chronology in Strupskardet. Former sea levels during the Lateglacial and Holocene are indicated with dotted lines in the lower part of the valley towards Lenangen. The former sea level at 87 m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

Moraines M10-M12

The ridges M10, M11 and M12 are located more or less continuously along the northeastern shore of Blåvatn (Fig. 4), whereas they are missing along the talus-covered southwestern shore. M10 is the most distinct ridge and can be traced from 30 to 3 m above Blåvatn for about 200 m. M11 and M12 are somewhat shorter, but like M10, they vary in height from 0.5-2.0 m and consist of unsorted sediments. The distance between the ridges is about 50 m, and the lower part of M12 disappears into Blåvatn c. 40 m beyond the modern glaciofluvial delta in the upper end of the lake. Lichen diameters on the three ridges are larger than 200 mm.

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

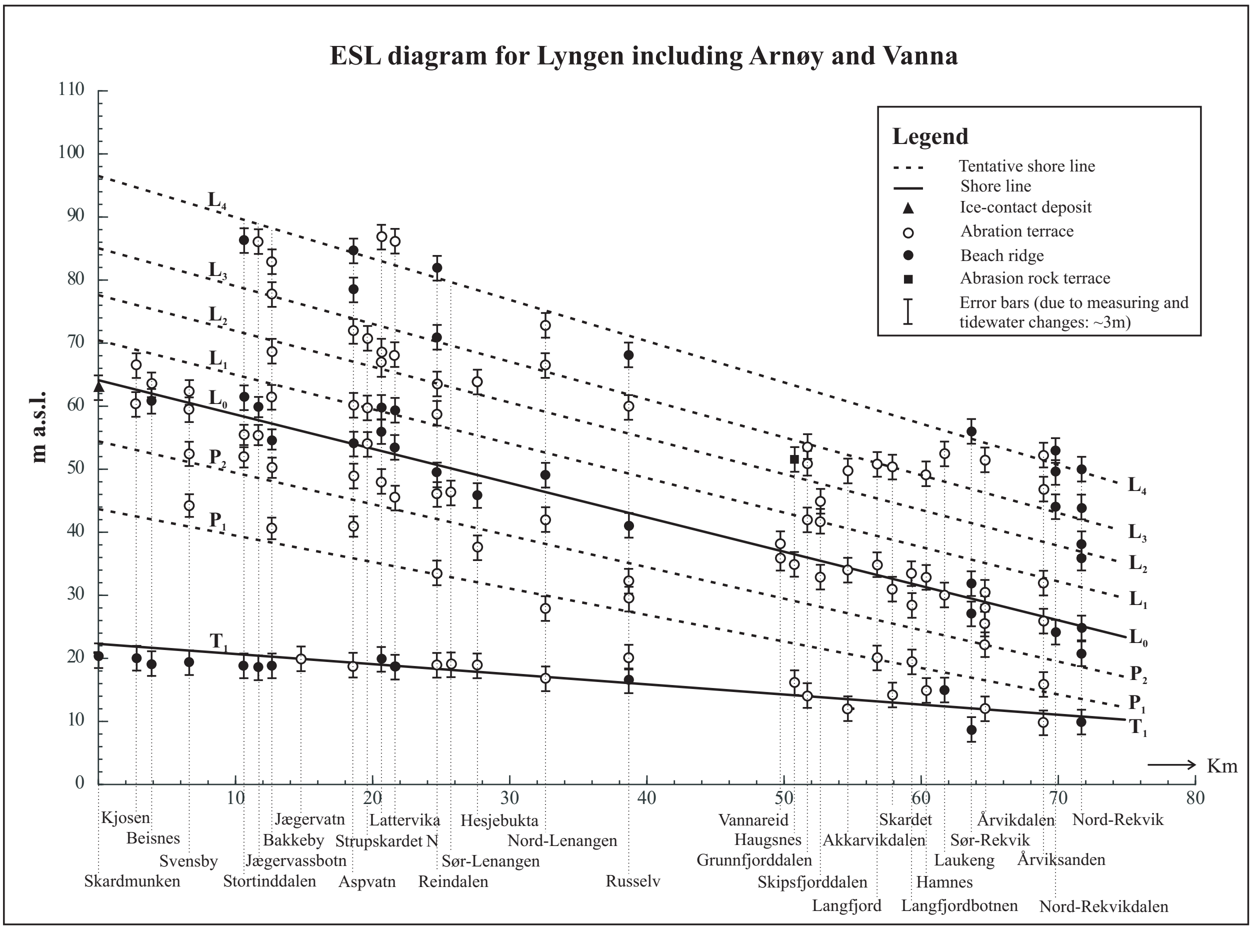

Figure 7 Equidistant shore-line (ESL) diagram for Lyngen, Arnøy and Vanna. The measurements from Russelv to Skardmunken are from to this study (82 measurements), whereas measurements from Arnøy and Vanna are adopted from Corner and Haugane (1993) and Kverndal and Sollid (1993). The most prominent former sea levels are the L_0 line / Main shoreline from the Younger Dryas and T_1 line from the Tapes transgression.

Moraine M13

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

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

Reconstructed glacier events

Based on the thirteen marginal moraines, reconstructed glacier events from Strupskardet are shown in Figure 6. They can be divided into 4 main stages:
- M1 to M3 are ice-cored moraines.
- M4 to M7 are push and melt-out moraines.
- M8 to M9 are likely ice-cored moraines.
- M10 to M13 are push and / or melt-out moraines.

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

Lateglacial and early-Holocene sea-level history in Strupskardet

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

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

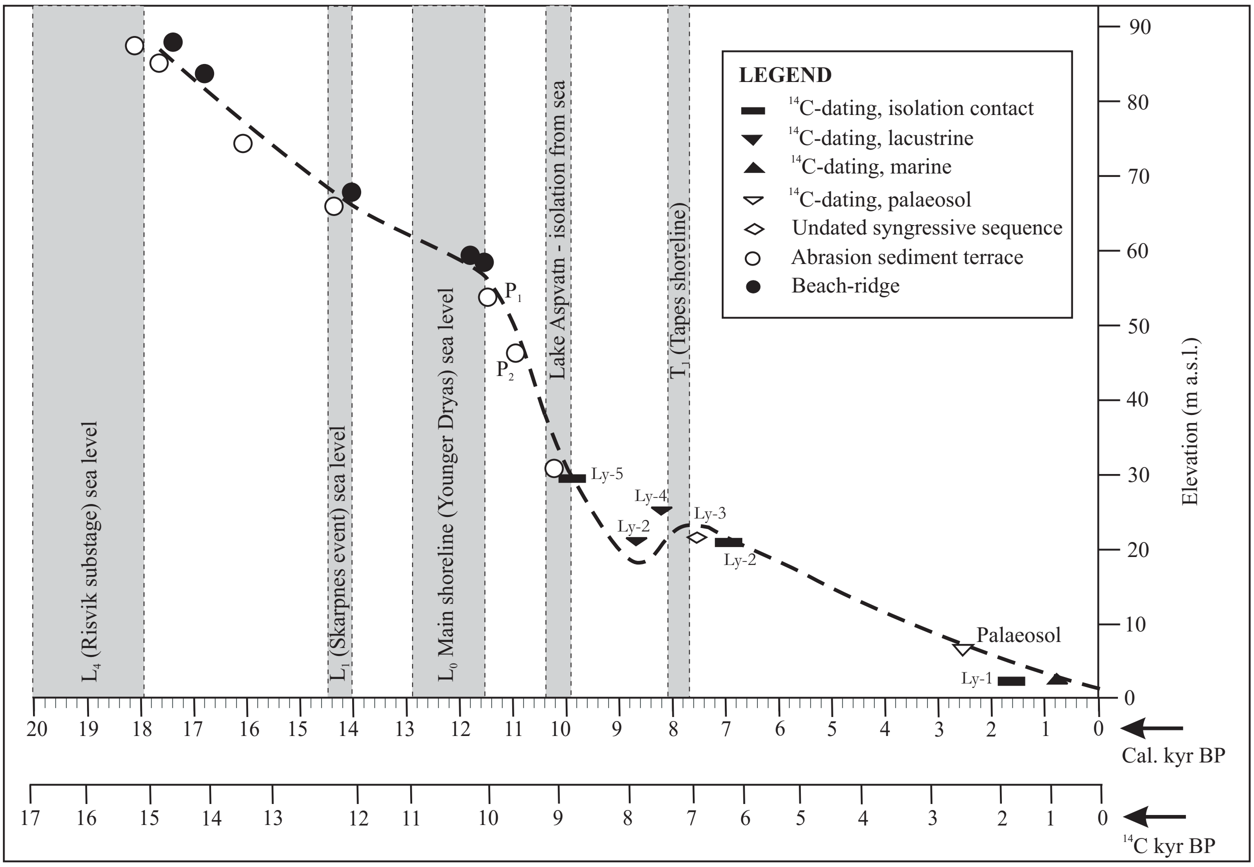

Figure 8 Shoreline-displacement curve from Strupskardet modified from Corner and Haugane (1993). The Holocene part of the sea-level curve is based on coring of isolation basins, whereas the Lateglacial sea-level curve is constructed based on the ESL diagram in Figure 7. Radiocarbon dates of shells from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

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

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

Holocene sea-level changes are based on the shoreline displacement curve constructed by Corner and Haugane (1993) from sites close to the study area in Strupskardet. As shown in Figure 6, the L_0 line/ Main shoreline corresponds to the Younger Dryas sea level, T_1 to the Tapes line, whereas L_1 – L_4 are older than the L_0 line. P_1 and P_2 are lines representing Preboreal sea levels. The L_4 line on Arnøy has been extrapolated inland to Jægervassbotn (Figs. 2 and 7), and is correlated to the Risvik substage with an estimated age of c. 20,000-18,000 cal. yr BP in Porsangerfjorden in Finnmark (Sollid et al., 1973; Andersen, 1979). At Svensby (Figs. 2 and 7), the L_1 line is suggested to represent the sea level during the Skarpnes event (c. 14,500-14,000 cal. yr BP) (e.g. Vorren and Plassen, 2002). Between the L_4 and L_0 lines in Strupskardet, however, no marked shorelines have been found, and the glacier in Ullsfjorden is therefore suggested to have retreated without any major halts or readvances at least to Svensby or Skardmunken (Paasche et al., submitted). As a consequence, the outer parts of Strupskardet may have been ice-free during at least the last 20,000 cal. yr BP.

Raised shorelines in Strupskardet

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

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

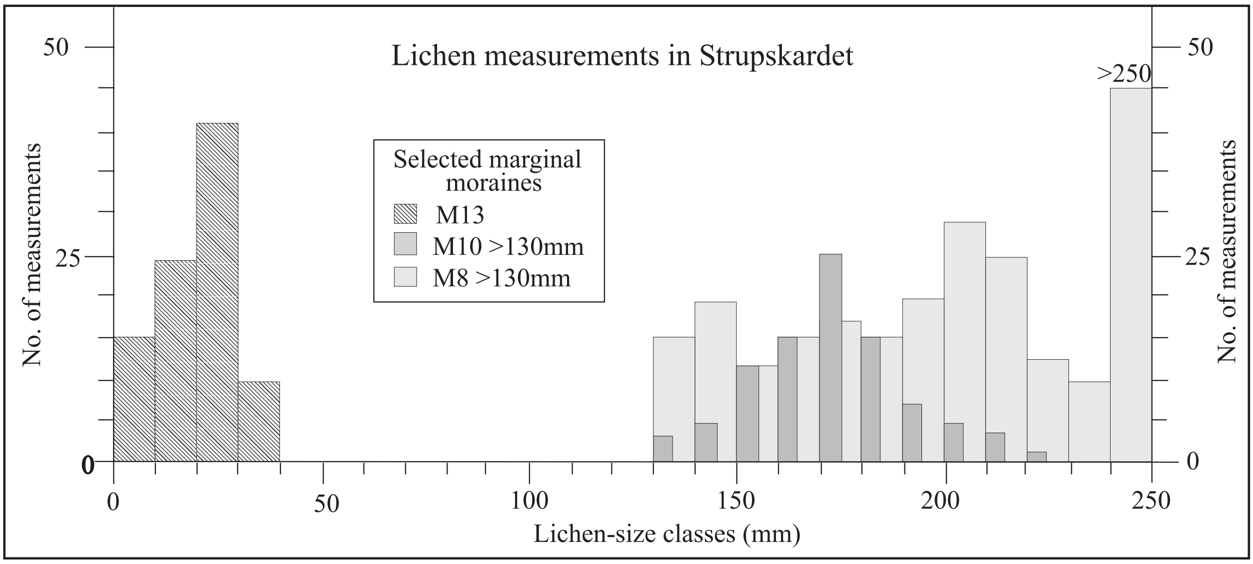

Figure 9 Lichen-size measurements on *Rhizocarpon geographicum* at selected moraines in Strupskardet. The M13 moraine contains lichens with maximum diameter of 40 mm, indicating an age close to the 'Little Ice Age' glacier maximum in Lyngen (e.g. Ballantyne, 1990). Similar measurements on the M10 and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

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

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

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

Distribution of relict periglacial features

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

Rock glaciers

Along the south-southwest facing valley side of Strupskardet, a series of 12 'hanging' lobes, mainly consisting of blocks and boulders with distinct fronts towards the valley, can be mapped for about 3 km (Fig. 4). The front of individual lobes are up to 10-30 m high with slope angles varying from 30-45°, and with a plateau behind consisting of transverse ridges with up to 5 m deep depressions in between. The distance from the back wall to the front of the lobe plateaux vary from 20 to 70 m, and the altitude of individual lobes drops from about 550 m in inner parts to 140 m towards the fjord. More-or-less active talus fans are present in the valley slopes uphill from the lobes, and several places these fans have filled in the lobe plateaux. In one case, a rather small lobe is found below another, but all described lobes are situated beyond the M1 moraine (Fig. 4).

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

Frost polygons

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

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

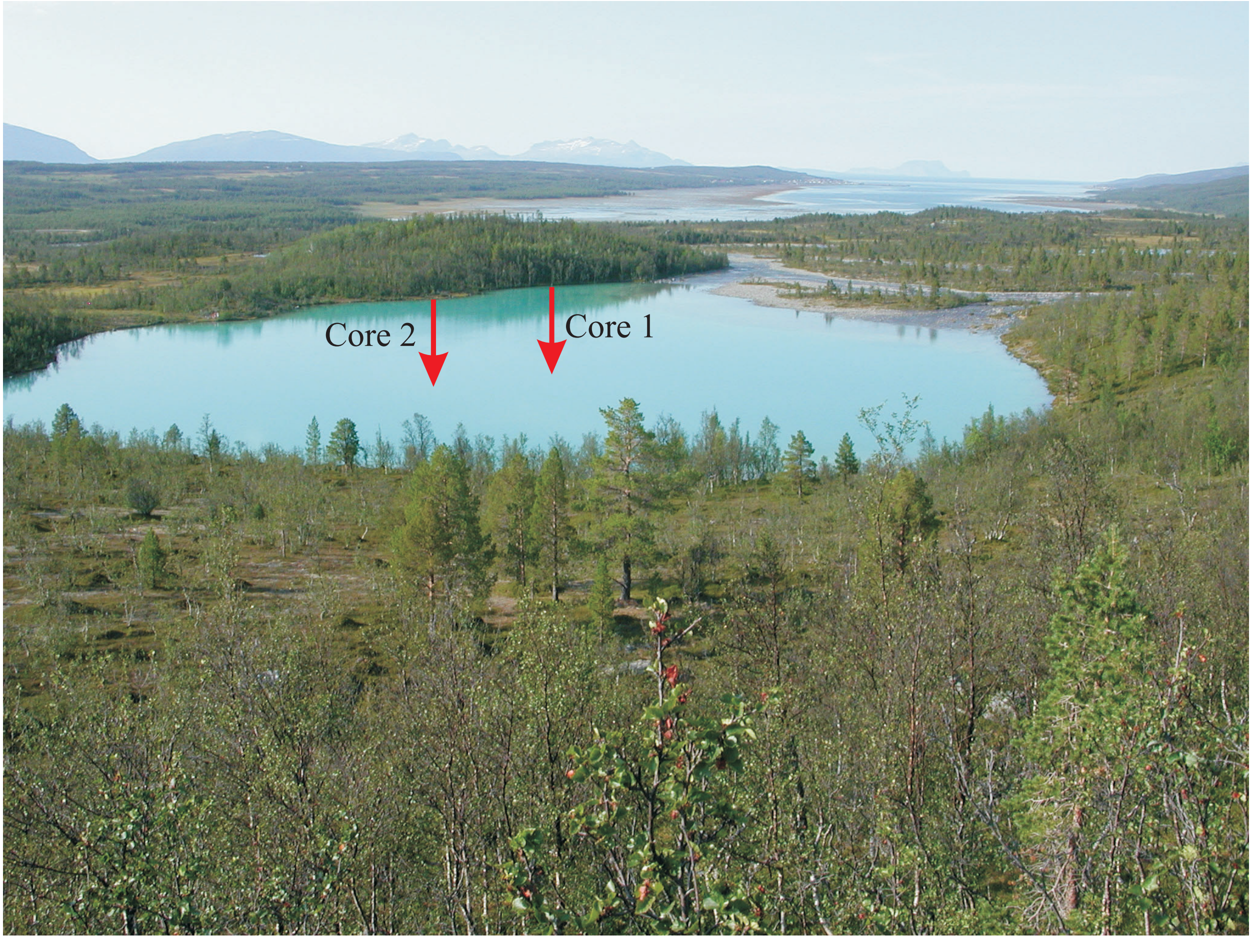

Figure 10 Aspvatnet with the inlet and outlet in the northern part of the lake. The fjord Lenangen in the background. Red arrows indicate positions of the two cores used in this study. Photo: Jostein Bakke.

polygons are suggested to be predominantly relict.

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

Lichenometry

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

The maximum diameter of 38 mm on M13 fits with the inferred 'Little Ice Age' maximum between A.D.1890 -1928 (e.g. Ballantyne, 1990), whereas the lichen diameters of M8 to M12 are well beyond the reliable range of modern growth curves in Norway (e.g. Innes, 1985; Matthews, 1994; Winkler et al., 2003).

Lacustrine sediments from lake Aspvatnet

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

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

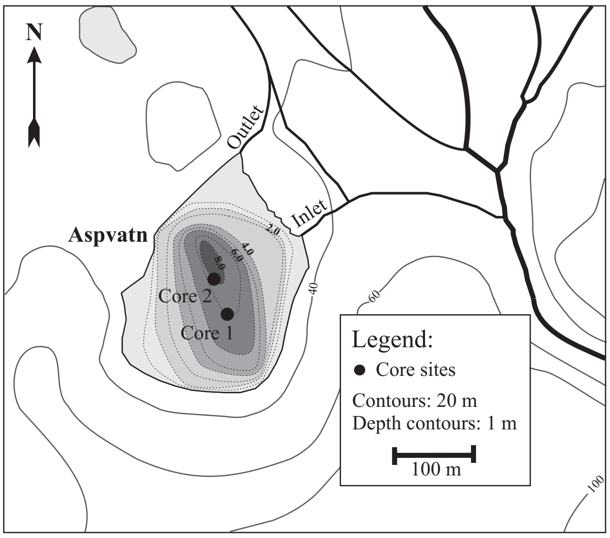

Figure 11 Bathymetric map of Aspvatnet. Thick lines show the main stream channels on the fan delta. Both the inlet and outlet in Aspvatnet are in the northern part of the lake.

will be dominated by coarser clastic sediments.

A distinct terrace 57 m a.s.l. around Aspvatnet is inferred to result from marine abrasion in the same period as a coarse fan-shaped delta along the present meltwater stream from Strupskardet was formed at this altitude. The fan-shaped delta was formed when the meltwater stream cut through M3, and the resulting gully appears to have been stable since because the modern stream lack competence to transport blocks up to 5 m in diameter present on the fan. Hence, the meltwater stream from Strupskardet is suggested to have followed this course during the Holocene. Only a small portion of the glacier meltwater entering this fan, however, ends up in Aspvatnet (Figs. 4 and 10).

Lithostratigraphy and radiocarbon dating

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

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

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

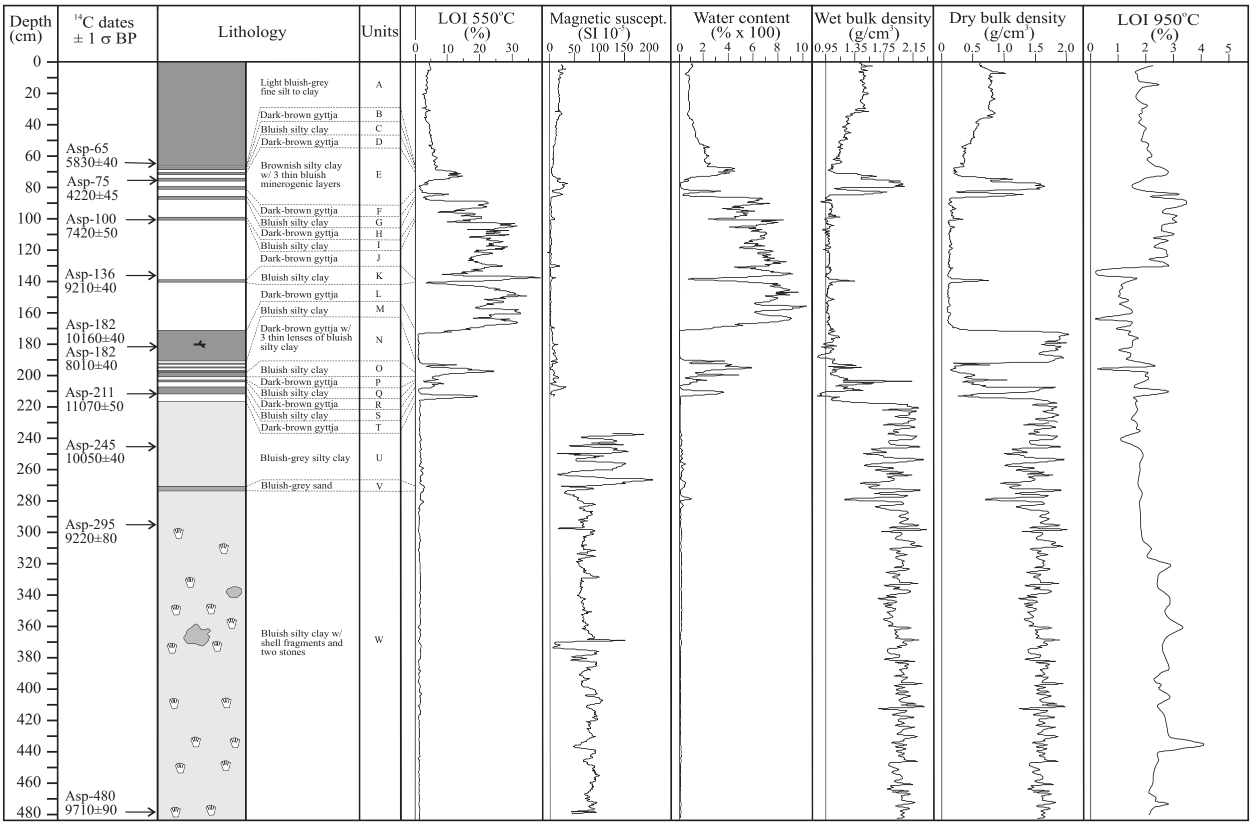

Figure 12 AMS-radiocarbon dated lithostratigraphy from core 1 in Aspvatnet with curves describing variations in loss-on-ignition (LOI) at 550˚C, magnetic susceptibility (MS), water content (in % of dry weight), wet bulk density (g/cm^3), dry bulk density (g/cm^3) and LOI at 950˚C. Because of problems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

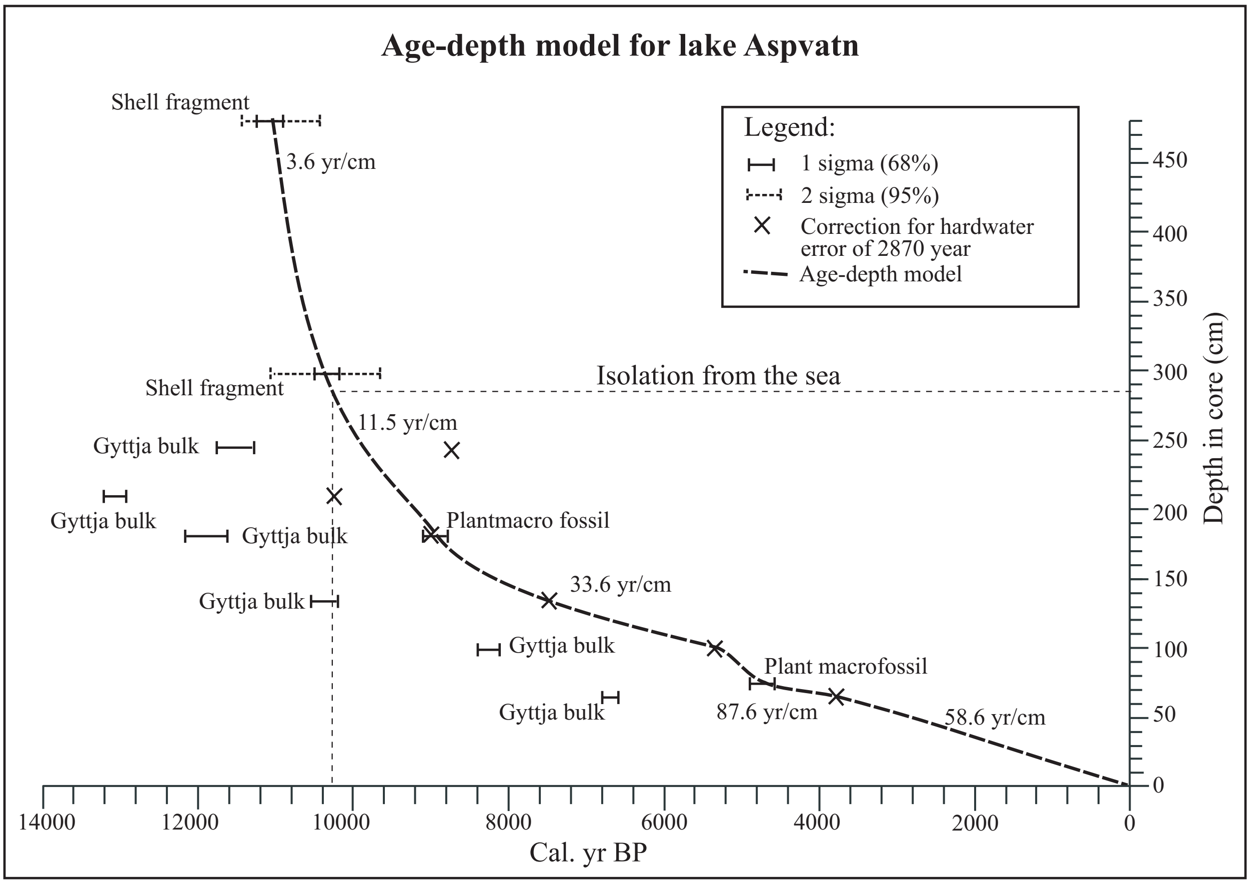

Figure 13 Age-depth model for core 1 in Aspvatnet based on the obtained AMS radiocarbon dates shown in Table 1. The two lowermost radiocarbon dates are obtained from marine shells and are corrected for a marine reservoir age of 400 ^{14}C yrs.

Discussion

The age of the marginal moraines in Strupskardet

The ice-cored moraines M1-M3 deposited by polythermal glaciers and the more-or-less contemporaneous talus-derived rock glaciers indicate a continuous permafrost regime in Strupskardet during the formation. A sub-marine formation of M1-M3 is regarded unlikely, and the corresponding sea level must therefore have been lower than 20 m for the ice-cored moraine complex to develop. Such a relatively low global sea level is inferred to have existed during the time span of maximum land-based ice volume from 22,000 to 19,000 cal. yr BP (Yokoyama et al., 2000). The L_4 sea level correlated to the Risvik substage (c. 20,000 – 18,000 cal. yr BP) in Porsangerfjorden in Finnmark (Sollid et al., 1973; Andersen, 1979) have abraded M1-M3. Hence, the age of these moraines must be somewhat older or from this time span (Figs. 4, 7 and 8).

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

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

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

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

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

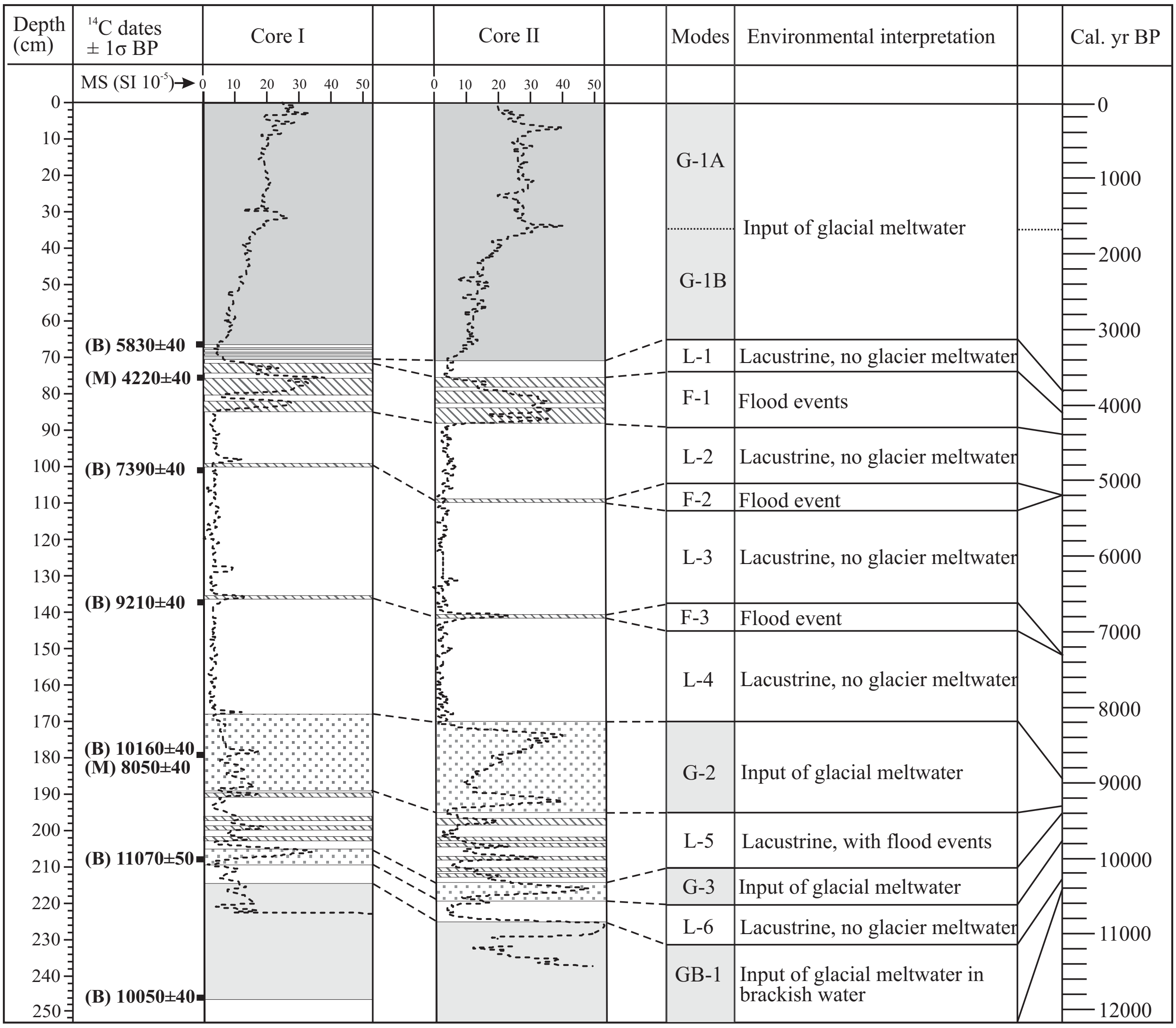

Figure 14 Composite log showing the lithology and environmental interpretation based on cores 1 and 2 in Aspvatnet. Four periods with input of glacier-meltwater induced sediments have been inferred. Three river floods have been identified in the lake record. There is good correlation between the two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

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

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

Sediment analyses from Aspvatnet

Adjusted for a marine reservoir age of 400 years, the sedimentation in Aspvatnet started around 11,100 cal. yr BP according to the basal date on shell fragments (Table 1). Hence, Aspvatnet can not be regarded as a lacustrine site before the isolation from the sea around 10,300 cal. yr BP (Fig. 8) (Corner and Haugane, 1993). The marine phase was characterised by high sedimentation rates (mean 3.6 yr / cm), and it is assumed that there was input of glacial meltwater to Lenangen (unit W in Fig. 12). Unit V marks the first isolation from the sea with a layer of fine sand, and the start of a period with brackish water and continuous input of glacier-induced sediments (unit U in Fig. 12, mode GB-1 in Fig. 14). Unit U culminated by a final isolation from the sea and transition to lacustrine conditions with no glaciers in the catchment in unit T (Figure 12, mode L-6 in Figure 14).

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

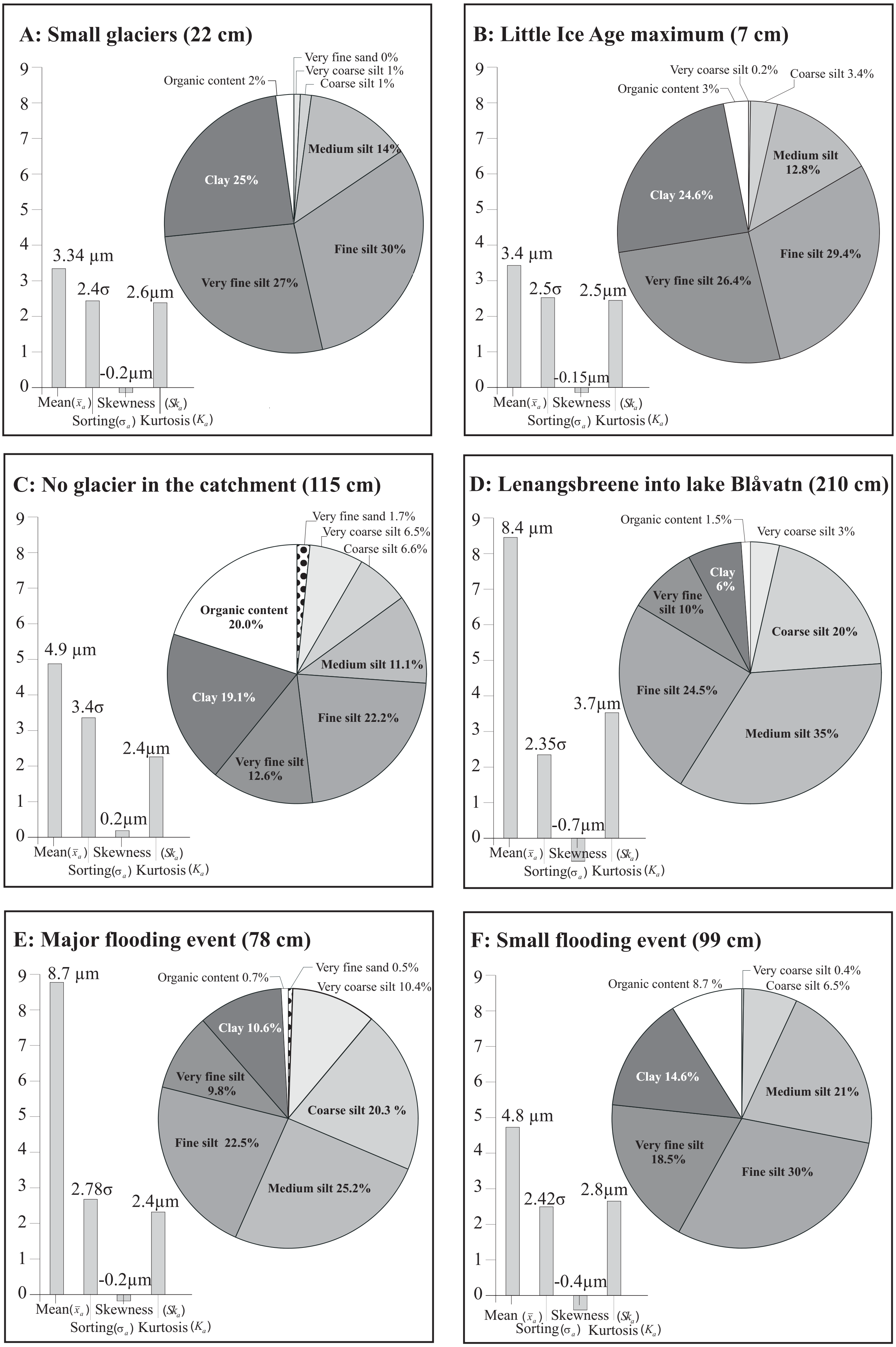

Figure 15 Six selected samples from core 1 in Aspvatn showing grain-size distribution, mean, sorting, skewness and kurtosis. Sediment sample A is from mode G-1A (small glacier), B is from mode G-1B (large glacier), C is from mode L-3 dominated by gyttja, D is from mode G-3 dominated by glacially derived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

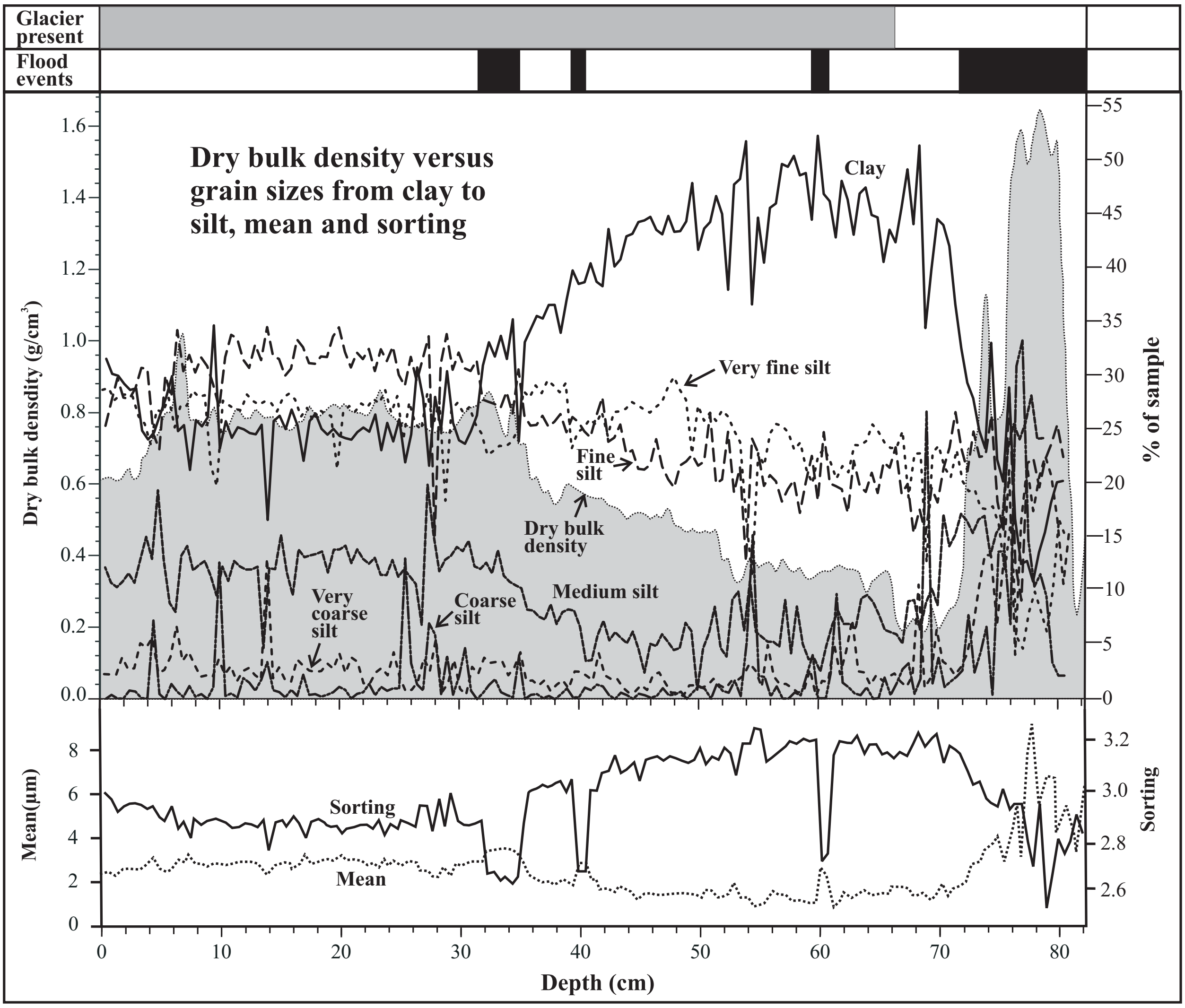

Figure 16 In the upper panel, the dry bulk density (DBD) from 0-82 cm depth in core 1 from Aspvatnet is shown as a grey-shaded area, whereas the grain-size distribution of clay and various silt fractions are presented in per cent of each sample. In the lower panel 'sorting' and 'mean' based on the grain-size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

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

Dry bulk density and grain-size analysis

Glacial erosion supplies bedrock particles to glacier-meltwater streams, and analysis of particle size may be a diagnostic parameter for glacier size (Matthews et al., 2000; Lie et al., in press). Wet-based, temperate glaciers produce abundant clay-silt size fractions that are transported downstream to produce characteristic signatures in glacio-lacustrine sediments (e.g. Østrem, 1975). The use of grain-size variations have, however, not been widely used in this context. One problem with the parameters is that it is difficult to flux correlate the signal, as the relatively long time intervals between the radiocarbon dates are a limiting factor. The distribution of grain sizes in proglacial lake sediments is mainly an indicator of the energy of the water masses flowing through the lake. As the sedimentation is closely related to Hjulström's diagram (Sundborg, 1956) of transportation in fluvial systems, it is apparent that high energy gives less of the finest grain sizes, and *vice versa*.

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

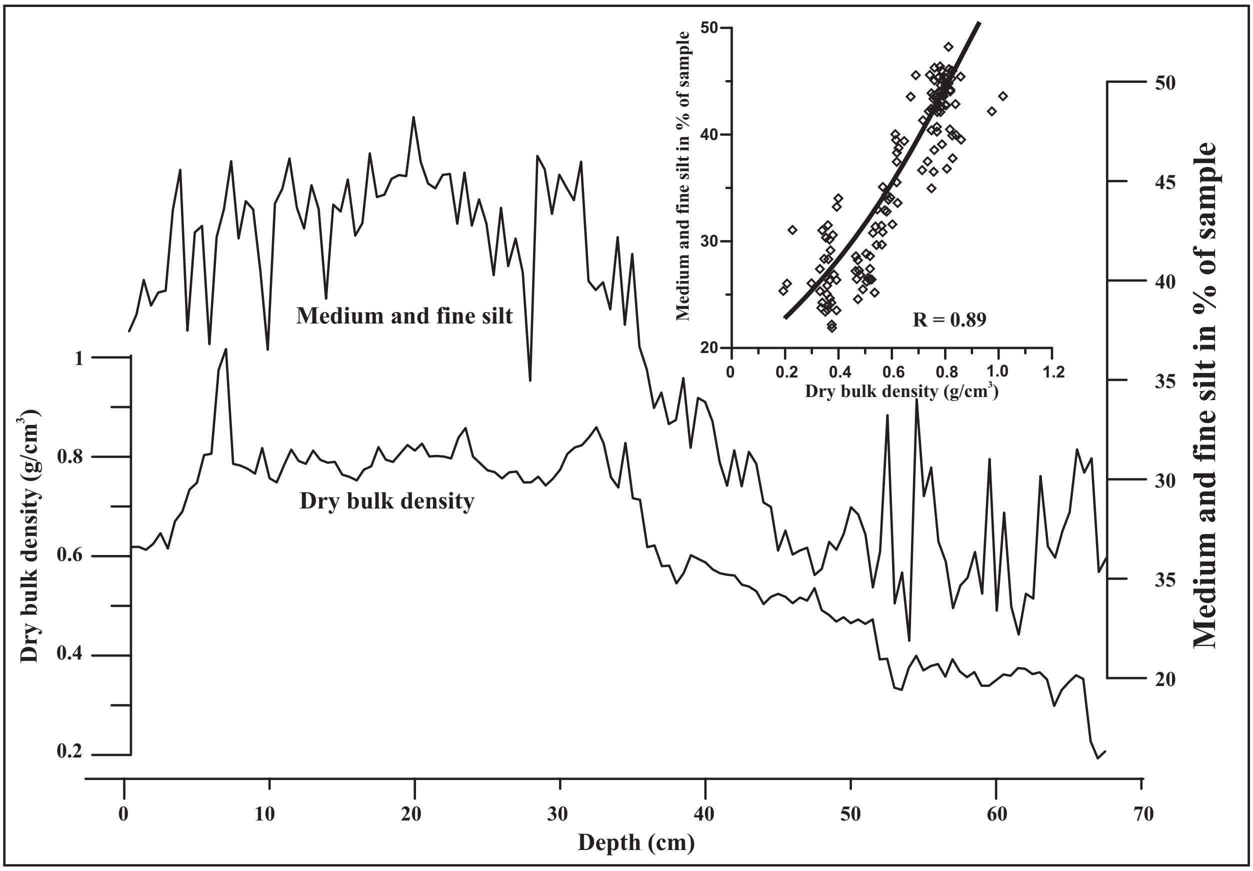

Figure 17 The main figure shows variations in a combination of fine and medium silt compared with fluctuations in dry bulk density (DBD) in the upper part of core 1 from Aspvatnet, whereas the incised figure shows the correlation (r=0.89) between dry bulk density and a combination of fine- and medium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

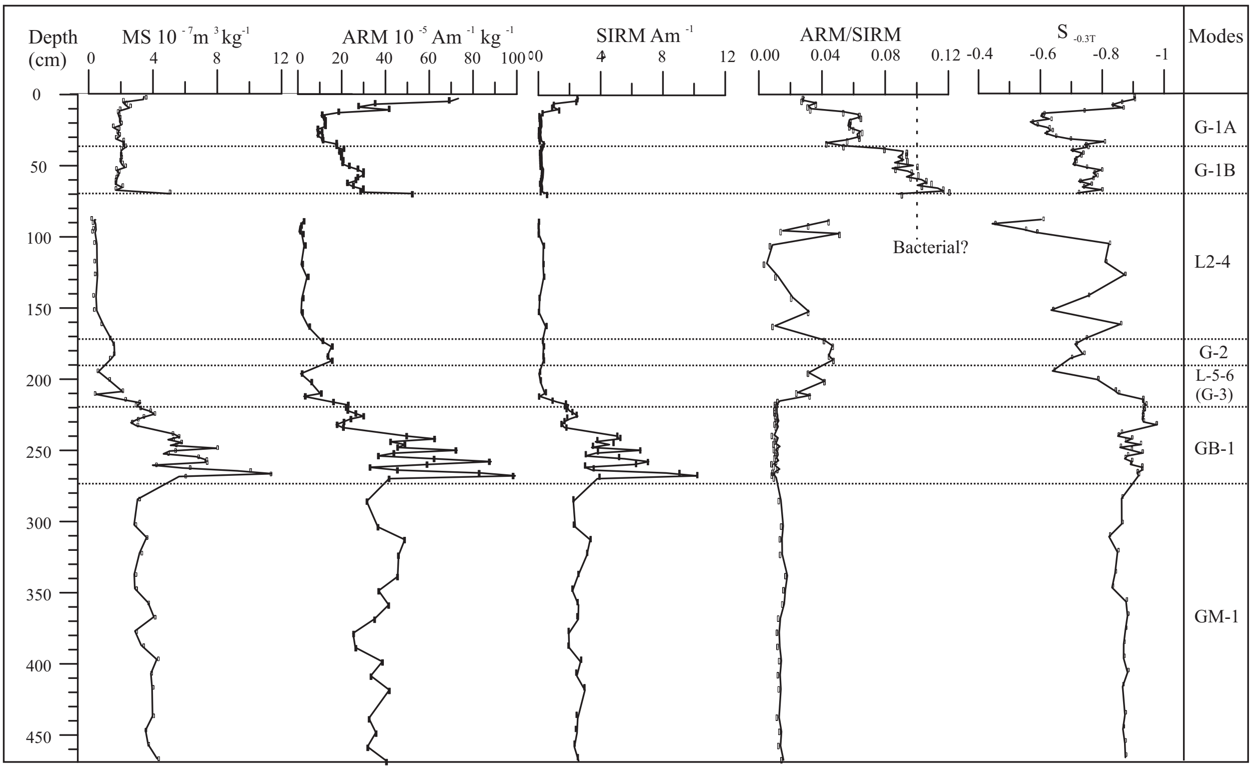

Figure 18 Diagram shows variations in environmental magnetic parameters in core 1 from Aspvatnet. Note the section with possible magnetotactic bacteria in the lower part of mode G-1B. Sample intervals vary with depth in the core. See text for further discussion.

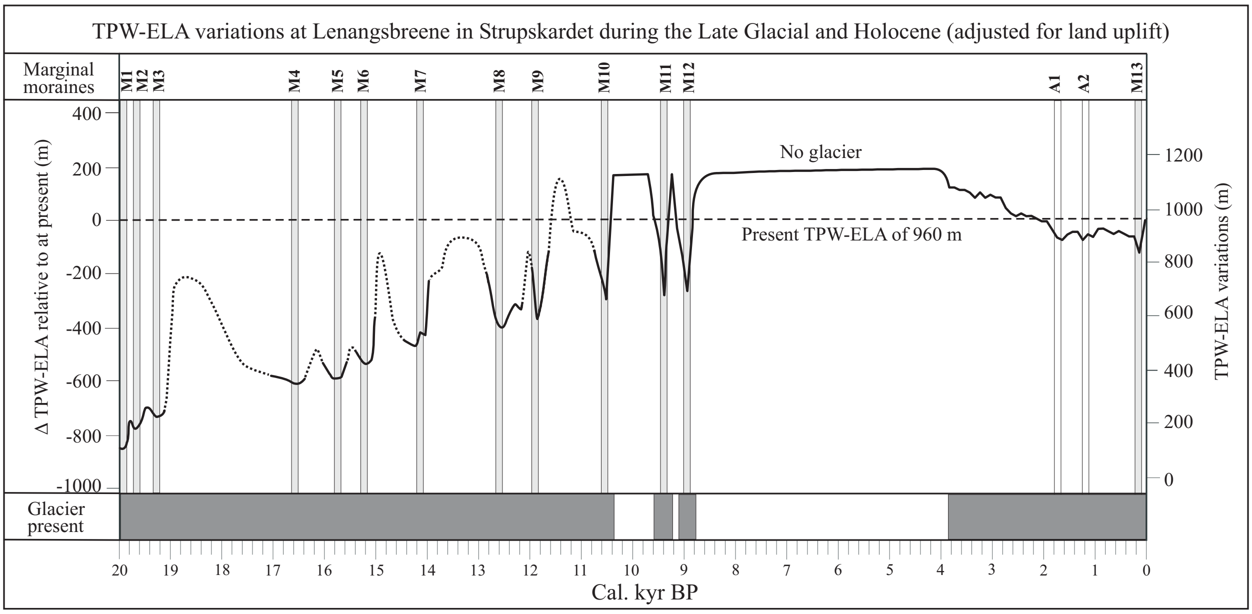

Figure 19 Mean TPW-ELA of Lenangsbreene in Strupskardet from 20,000 cal. yr BP to at present. The TPW-ELAs are reconstructed using an AAR of 0.6 on the moraines M8 and M13 and a technique termed the 'Little Ice Age ratio' (LR) (Dahl *et al*., 2002) adapted to Strupskardet for the remaining marginal moraines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

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

During the early-Holocene glacier events that formed the moraines M10-M12 (Fig. 4), sedimentation in Aspvatnet was dominated by group 4 (mode GB-1, G-3 and-G-2) and group 3 (mode L-6, L-5 and L-4) with abrupt changes between clastic sedimentation and gyttja (Figs. 14 and 15). However, two (three) episodes during the early- and mid-Holocene are recognised as small low-energy river floods belonging to group 6 (mode F-3 and F-2), and a major flooding event, representing group 5 (mode F-1), occurred from 85-73 cm in core 1 (Figs. 12 and 14). Group 1 is recognised as small glaciers (mode G-1B), whereas group 2 represents large glaciers ('Little Ice Age' maximum) (mode G-1A) in the upper part of core 1 linked to moraine M-13 (Figs. 4, 12, 14 and 15).

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

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

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

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

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

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

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

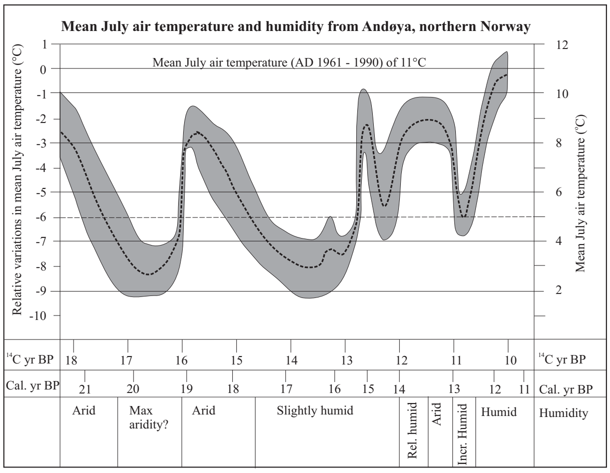

Figure 20 Reconstructed mean July air temperature and humidity at Andøya, northern Norway. The dotted line is used as temperature estimate during the Lateglacial for quantification of former winter precipitation in Lyngen. Modified from Vorren *et al.* (1988).

Magnetic modes

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

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

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

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

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

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

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

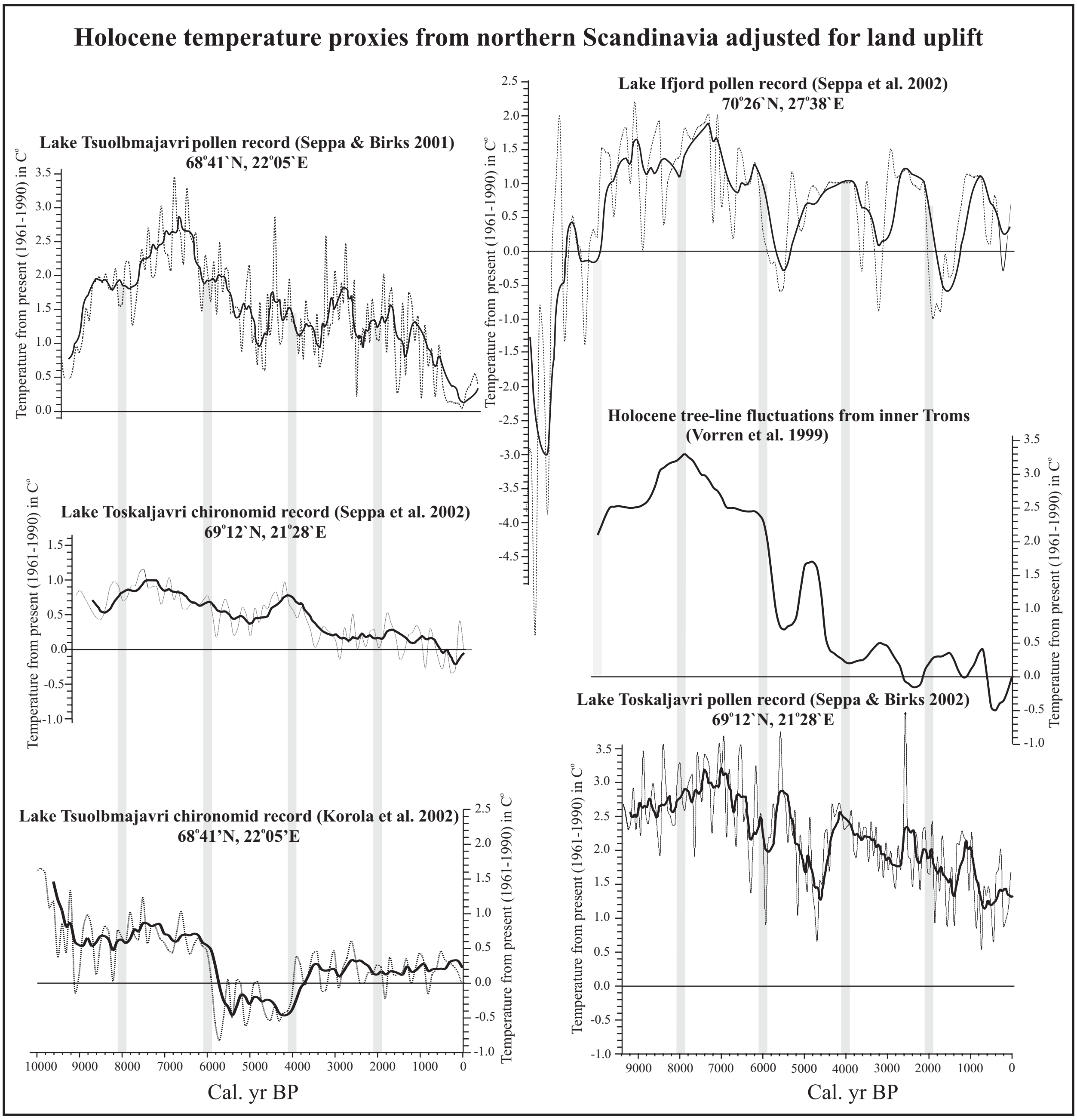

Figure 21 Five selected continuous temperature reconstructions using biological proxies from lakes and one using pine-tree limit fluctuations in northern Scandinavia are adjusted for land uplift. The reconstructions, taken into account vegetation changes from the lakes Tsuolbmajavri (Seppä and Birks,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

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

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

- G-1A (32-12 cm, N=11) represents a phase with small glaciers in the catchment, and has a constant mixture of SD and larger grains. ARM / SIRM is significantly higher than below mode L-2, L-3 and L-4. Since an ARM / SIRM ratio less than 0.1 does not exclude the presence of bacterial magnetite, magnetosomes may be present in this zone, but are diluted with larger, detrital magnetite grains. With S-ratios significantly below –1.0, the occurrence of bacterial magnetite is uncertain.

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

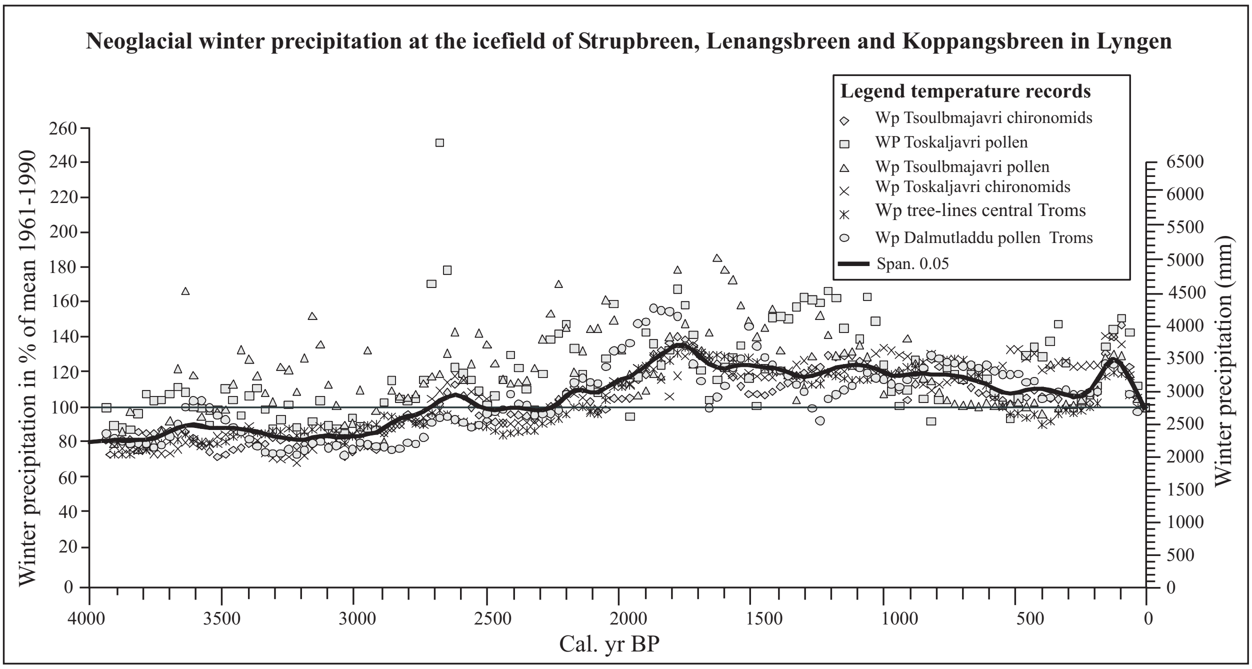

Figure 22 Mean winter precipitation at Lenangsbreene from 3800 cal. yr BP to at present. The thick line is a lowness filter with a span of 0.05 which gives the average winter precipitation estimate based on calculations using the different temperature reconstructions in Figure 21. In addition, an unpublished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

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

Lateglacial and Holocene variations in magnitude and TPW-ELA

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

The M8 glacier is well-defined by use of the marginal moraines, and the M13 glacier is well-defined based on lichenometry and local topography in front of Lenangsbreene (Figs. 4 and 6). The reconstructed glacier profiles of M8 and M13 are relatively steep and they are both suited for the use of AAR to calculate former TPW-ELAs. By use of an AAR of 0.6, the former TPW-ELAs of the M8 and M13 glaciers were -330 and -120 m, respectively. The marginal moraines beyond M8, including M1-M3, are located on relatively flat ground and thus less suited for use of AAR (e.g. Nesje and Dahl, 1991). A new technique termed the ‘Little Ice Age ratio’ (LR) (Dahl et al., 2002) adapted to Strupskardet has therefore been used to estimate former TPW-ELAs on the remaining marginal moraines. The distance between the termini of M13 and M8 is 1500 m and corresponds to a drop in the TPW-ELA of 210 m. By using this ratio (ELA lowering (m) / distance (m)) not adjusted for land uplift, corresponding TPW-ELAs were estimated on the remaining marginal moraines, and shown in Table 2 and Figure 19 adjusted for land uplift.

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

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

The relict ice-cored moraine complex (M1-M3) in Strupskardet is assumed to have developed while the sea level was lower than 20 m a.s.l during the time span of maximum global land-based ice volume from 22,000 to 19,000 cal. yr BP (Yokoyama et al., 2000), and before this complex was abraded by the L_4 sea level during the Risvik substage no later than c. 20,000 - 18,000 cal. yr BP. This is also comparable with the glaciation diagram from Andfjord-Vågsfjord, where the glacial events Bjerka and Egga II are placed withBennin this time interval (Vorren and Plassen, 2002). The rapid global decrease in land-based ice volume after 19,000 cal. yr BP (Yokoyama et al., 2000) with release of water to the sea, fits well with the assumed age of the L_4 sea level in Lyngen with an estimated altitude of c. 87 m in Strupskardet (Figs. 7 and 8).

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

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

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

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

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

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

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

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

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

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

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

After 3800 cal. yr BP, there has been a continuous input of glacier-meltwater derived sediments to Aspvatnet. To obtain a continuous TPW-ELA record for this time span, a calibrated dry bulk density (DBD) curve (Fig. 17) has been used. Major glacier events are suggested to have taken place c. 1800 and 1200 cal. yr BP (A1 and A2 on Figure 19), but moraines from these episodes were probably erased by the maximum 'Little Ice Age' glacier c. AD 1910. Based on an AAR of 0.6, the TPW-ELA of the 'Little Ice Age' maximum M13 glacier was c. 120 m, whereas the suggested TPW-ELAs for A1 and A2 were lowered by about 80-40 m compared to modern values (Table 2, Fig. 19). From southern Lyngen the ELAs lowering during the 'Little Ice Age' (AD 1880-1910) were calculated to 100-250 m compared to modern values (Gellatly, 1988; Ballantyne, 1990).

Temperature reconstructions from northern Scandinavia

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

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

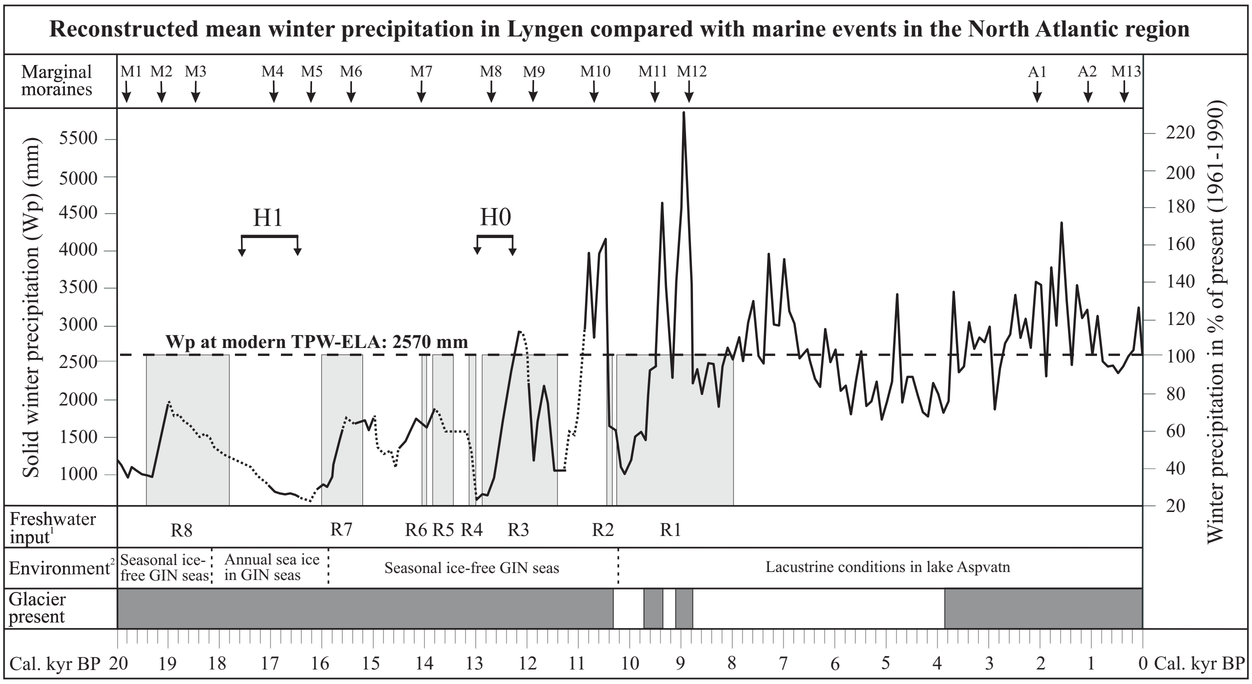

Figure 23 Variations in mean winter precipitation from 20,000 cal. yr BP to at present based on reconstructed TPW-ELAs from Strupskardet in Lyngen, Lateglacial temperature estimates from Andøya (Vorren *et al*., 1988), early Holocene (11,500 to 10,500 cal. yr BP) temperature estimates from Ifjord (Seppä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

Lateglacial and Holocene winter precipitation variations in Lyngen

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

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

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

Lenangsbreene consist of two nearly coalesced cirque glaciers which receive a large amount of windblown snow compared to the regional mean in this part of Lyngen. Hence, the ELA of Lenangsbreene reflects a local TPW-ELA. As Lenangsbreene is part of a larger ice field, including glaciers in most aspects, however, the combined TPW-ELAs of these glaciers may be regarded as representative for a regional TP-ELA in Lyngen (Fig. 5). Based on the inferred 'Little Ice Age' and Younger Dryas moraines in Figure 5, this is even more likely when these glaciers are in a more advanced position than at present.

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

Comparison with Holocene glacier records in northern Scandinavia

The Holocene glacial history obtained in this study deviates in some periods with that suggested for northern Sweden (Karlén, 1973; Karlén and Denton, 1975; Karlén, 1976) and for northern Norway (Griffey and Worsley, 1978; Ballantyne, 1990; Winkler et al., 2003). However, the inferred 'Little Ice Age' maximum c. A.D. 1900 at Lenangsbreene and the lacustrine record from Aspvatnet, indicating that the glacier was almost as large c. A.D. 1800, fit well with previous studies from northern Scandinavia. At some sites at Svartisen, Okstindan, Saltfjellet and Kebnekaise, the late- Holocene glacier maximum may have occurred earlier, with advances dated at A.D. 1750, 1810, 1850 and 1890 (Karlén, 1976).

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

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

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

Comparison with glacier records from southern Norway

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

- The degree of continentality with different responses to changes in summer temperature and winter precipitation.
- Morphological differences related to altitude, aspect, gradients and magnitude.
- Different atmospheric circulation patterns during for instance the 'Little Ice Age'.

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

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

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

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

Comparison with marine climate archives in the North Atlantic region

To put the inferred Lateglacial and Holocene climate from Lyngen in a North Atlantic context, the winter-precipitation record from Strupskardet has been compared with the timing of Heinrich events (Clark et al., 2002), melt-water rerouting events from the Laurentide ice sheet (Clark et al., 2001) and the sea-ice record based on marine diatoms (Koc et al., 1993) (Fig. 23). The mean Lateglacial winter precipitation in Lyngen was c. 50-60 % compared to at present and seasonal sources of precipitation is suggested to have existed. Open water in the Norwegian Sea, however, assumes production of North Atlantic Deep Water (NADW), and according to Dokken and Jansen (1999) this may have taken place along the western coast of Norway because of brine formation during Weichselian stadials.

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

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

Main Conclusions

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

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

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

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

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

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

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

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

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

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

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

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

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

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