Holocene millennial-scale summer temperature variability inferred from sediment parameters in a non-glacial mountain lake: Danntjørn, Jotunheimen, central southern Norway

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Abstract

Holocene changes in summer temperature in the Jotunheimen area have been inferred from variations in several sediment parameters from a composite (combined from two overlapping sediment cores) lake-core record (mean time resolution of \sim15 years) from Lake Danntjørn, a small mountain lake located close to the modern pine-tree limit in Jotunheimen, central southern Norway. Loss-on-ignition analysis at contiguous 0.5-cm intervals throughout the cores revealed that the variations in residue after ignition of the sediments at 550 °C in the upper 260 cm (~8500 calendar years) were not associated with variations in magnetic susceptibility. In the basal, deglacial and early Holocene sequence of the lake sediments, however, the two sediment parameters fluctuated mainly in phase. In addition, the biogenic silica (BSi) concentration is relatively high. Thus, in the early part of the Holocene, the variations in the residue were probably primarily driven by input of minerogenic, detrital material (inwash and/or windblown) and biological productivity (including BSi from diatoms) in the lake itself and in the lake surroundings. From approximately 8500 calendar years BP, variations in the residue probably primarily reflect variations in low-minerogenic (low magnetic susceptibility with little variations) and diatom (BSi) productivity, as demonstrated with analysis of BSi concentration between 260 and 233 cm (~8500–7760 calendar years BP). The highest residue values (probably mainly reflecting high diatom productivity) occurred at 7800–7200, 7000, 6700, 6200, 5400–5050, 4300, 2500, 1800, 1550, 900, 700, and \textasciitilde100 calendar years BP. The most significant residue minima (mainly low diatom productivity) were recorded at 7100, 6800, 6600, 5000–4400, 4200, 3800, 3600, 3200, 3000, 2750, 2300, 1900, 1700, 1070, 740, 600, 380, and 80 calendar years BP (BP = AD 2000). The lake record from Jotunheimen is discussed in the context of Holocene terrestrial and marine records from the North Atlantic region and possible natural climate forcing factors.

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1. Introduction

Continental climate records from Scandinavia capture Holocene insolation changes as well as changes in North Atlantic Ocean thermohaline and atmospheric circulation. In order to make reliable and likely predictions of future climate, and to separate natural from human-induced climate variability, it is important to know the rate and magnitude of past climate changes (e.g. Alverson et al., 2003). Because instrumental meteorological records commonly are too short to cover the entire climate variability, the climate history for earlier periods has to be reconstructed from indirect (proxy) indicators. Previously, climate reconstructions from biological sedimentary remains were based on single indicator species or assemblages of taxa and the results were usually interpreted in a qualitative and descriptive manner. Numerical techniques and approaches are now available that allow quantitative reconstructions...
from floral and faunal assemblages (ter Braak and Juggins, 1993; Birks, 1995, 1998, 2003). Holocene climate variations in Europe and the north Atlantic region have been reconstructed from a number of proxies and archives (Table 1).

Mountain lakes respond quickly to the environmental impact of climate changes. Lakes close to the modern (or past) tree lines or at ecotone boundaries are commonly more sensitive to such changes than lowland lakes (e.g. Körner, 1998). Such lakes are therefore used as sensors of modern and past environmental changes and climate variations. Climate variability observed from instrumental records (e.g. Schindler et al., 1996) over the last few decades has been recorded as changes in lake ecosystems inferred from sediment records. Climate variations influence lakes in different ways and the direct and indirect linkages between climate and lake sediments need to be understood in order to realise the potential of using lake sediments to obtain records of past climate variability. The key processes are those affecting radiation balance and water balance. Radiation determines light and temperature regimes that are modulated by winter lake ice and snow cover and by wind. These factors then influence depth, duration and intensity of water stratification affecting chemical and biological processes (primary production, nutrient cycling, oxygen consumption and pH).

Lake sediments commonly accumulate quite rapidly and sediment cores can be sub-sampled at mm-interval to provide data with decadal to sub-decadal time resolution. Climate reconstructions from lake sediments rely on relationships between the sediment record and the climate. The use of biological proxies may involve the use of transfer functions (e.g. Birks, 2003). Large diameter (~110 mm) cores together with non-destructive techniques, or techniques that require only small samples, make it possible to carry out multi-proxy studies. Climate reconstructions using a multi-proxy approach are becoming increasingly common (e.g. Ammann et al., 2000; Birks et al., 2000; Lotter and Birks, 2003). Reliable inter-core correlations may be carried out using simple physical sediment parameters (e.g. dry weight, water content, loss-on-ignition, bulk density, magnetic susceptibility).

Recent research on lake and marine cores has demonstrated that the climate of the Holocene has been more variable than previously recognised. Organisms in ecotonal environments may be responsive to climatic variations because they are close to their physiological limits of their distribution (Lotter and Birks, 2003). Most commonly, summer temperature, the length of the growing season, the depth/duration of the snow cover, and wind strength/exposure are the most controlling physiographic factors. It is only in the past few decades that lacustrine records from near ecotones have been the focus of detailed palaeoclimatological studies (e.g. Battarbee et al., 2002). Mountain lakes are particularly sensitive recorders of past and present climate change as the occurrence and composition of aquatic and

Table 1
Examples of Holocene climate reconstructions in Europe and the North Atlantic region from different proxies and archives


Stable-isotope records from tree rings: See review by McCarroll and Loader (2004).


terrestrial biota are directly or indirectly related to climate (e.g. Lotter et al., 1997; Battarbee, 2000). The length of lake ice-free period and the summer temperature are commonly the main controlling factors of the primary production in the water column in addition to the mixing regime and the amount of oxygen depletion (Livingstone, 1997). Catchment soils and vegetation have also a strong influence on water chemistry (e.g. Birks et al., 2000).

The main aim of this study was: (a) to reconstruct the Holocene environmental variability from several sediment parameters in a non-glacial mountain lake close to the modern pine-tree limit in Jotunheimen, central southern Norway, and (b) to discuss the inferred Holocene temperature variability record from Jotunheimen in the context of other terrestrial and marine records in the North Atlantic realm and with the respect to possible natural climate forcing factors.

2. Lakes as climate archive

2.1. Introduction

Commonly, lake sediments consist of three components (in varying proportion): organic matter, none-carbonate clastic material, and one or more carbonate minerals. In lake sediments, the organic content reflect autochthonous production from plants and input of eroded organic and minerogenic material from the catchment. The organic material in lake sediments may be produced in the lake itself, washed in from rivers in the catchment, and blown in from the lake surroundings. The minerogenic material in lakes may originate from lake-shore erosion and wave action, lake-ice erosion along the shallow shores, transport by glacier meltwater streams, fluvial erosion of deposits along the rivers in the catchment, redeposition of material by slumping and turbidity currents, and windblown material (local and regional sources). Slope processes, such as gullies, debris flows, and ‘wet’ snow avalanches may also contribute to minerogenic and/or organic input to lakes due to erosion of soils and clastic material in the catchment. In proglacial lakes, high variability of minerogenic input to the lacustrine system is generally higher than the internal variability/production of organic material. In such lakes, changes of the organic/minerogenic content mainly reflect allochthonous, glacially produced minerogenic influx. If the dominant process(es) of the minerogenic material input to a lake is (are) identified, specific lake catchment processes may be reconstructed in detail from lacustrine sediments. Mapping and monitoring of processes in the catchment are therefore recommended in order to better understand variations in the sediment parameters.

2.2. Organic matter

The organic matter content of lake sediments provides a variety of indicators that can be used to reconstruct palaeoenvironmental variations of lakes and their catchments. Organic matter originates from organic matter components (lipids, carbohydrates, proteins, etc.) produced by organisms that have lived in and around the lake (e.g. Meyers and Teranes, 2001). Organic matter derived from the residues of plants is gradually altered into humus through physical fragmentation, faunal and microfaunal interactions, mineralisation and other processes of humus formation. Measurement of the organic content of soils and sediments is either by loss-on-ignition (LOI) or determination of organic carbon (OC) content by the wet oxidation technique. The inorganic carbon fraction may consist of coal, charcoal, and carbonates. Total organic carbon (TOC) is commonly measured in the analysis of lake and marine sediment cores by means of an LECO carbon-carbonate determination.

LOI is the most commonly applied method for estimating the organic content in lake sediments (e.g. Dean, 1974; Håkanson and Jansson, 1983; Heiri et al., 2001 and references therein). LOI has been shown to be a remarkable ‘composite’ proxy of environmental change in high latitude lakes (Willemsen and Törnquist, 1999; Battarbee et al., 2001, 2002; Nesje and Dahl, 2001; Kaplan et al., 2002). The interpretation of the LOI signal may, however, be complex as the residue after ignition at 550 °C may consist of varying inputs of inorganic mineral matter, carbonate and biogenic silica (diatoms) (e.g., Battarbee et al., 2002). The LOI values are also influenced by varying input and preservation of organic matter. The LOI signal may also be controlled by sedimentation rate changes of mineral matter (most relevant in proglacial lakes). Both inorganic and organic matter can be produced within the lake as well as in the catchment. Small mountain lakes located on crystalline bedrock, with sparse vegetation and thin soils in the catchment, few or small inlets, and restricted aquatic macrophyte flora, are commonly well suited because the factors influencing on the LOI signal are fewer.

Differential thermal analysis (DTA) thermograms show that when a dried, powdered sample containing organic material and calcium carbonate is heated in a muffle furnace, the organic material begins to ignite at about 200 °C and is completely ignited by the time the furnace temperature has reached approximately 550 °C. The high correlation (r = >0.95) between ignition loss organic matter and percent organic carbon determined chromatographically shows that the LOI method is a measure of the amount of organic matter in a sample [the weight LOI is 2.13 ± 0.4 times the organic carbon content (Dean, 1974)]. Snowball and Sandgren (1996)
showed that LOI was 2–3% higher than TOC, probably due to crystalline water in clayey material.

According to Dean (1974) evolution of CO₂ from the calcium carbonate will begin at about 800 °C and proceed rapidly so that most of the CO₂ has been evolved by the time the furnace has reached 850 °C. If any dolomite is present in the sample, it will evolve CO₂ at a lower temperature than calcite (at approximately 700–750 °C). Sutherland (1998) suggested that structural water may be lost by metal oxides at temperatures as low as 280–400 °C. Most clays contain up to 5% lattice OH water which is not removed until heated to 550–1000 °C. Therefore, the 550–950 °C ignition loss would contain a significant amount of lattice water in samples that are low in carbonate and high in clay. For example, lake sediments containing no carbonate but with high clay content usually show a 2–4% LOI between 550 and 950 °C. Sediments containing 100% clay would presumably yield up to 5% weight loss between 550 and 950 °C.

Inorganic carbon may be lost at temperatures between 425 and 520 °C in minerals such as siderite, magnesite and rhodochrosite (Weliky et al., 1983; Sutherland, 1998). There is also a possible loss of volatile salts at 550 °C (Bengtsson and Enell, 1986). The LOI method does not, however, indicate which carbonate minerals that are present. Different carbonate minerals will evolve CO₂ at slightly different temperatures, but they cannot be separated by this method, thus giving total carbonate.

Diatoms are photosynthetic algae that secrete biogenic silica (SiO₂ nH₂O) as an internal shell. Biogenic silica (BSi) reflects the sedimentary abundance of diatoms, which are single-celled algae that commonly dominate lake primary productivity (Wetzel, 2001).

The isotopic composition of autogenic and biogenic carbonates and diatom silica are used as palaeoclimatic proxies from lake sediments (e.g. Leng and Marshall, 2004). The oxygen isotopic composition of BSi depends on the ambient water temperature and the isotopic composition of the lake water when the shell is secreted (e.g. Shemesh et al., 1992). In lacustrine environments stratigraphic changes in δ¹⁸O values are commonly attributed to changes in temperature or precipitation/evaporation ratio, whereas carbon and nitrogen isotopes are used to infer changes in carbon, nutrient cycling and productivity within lakes and their catchments.

2.3. Mineral matter

Magnetic susceptibility of lacustrine sediments is a useful indicator of erosion and transport of clastic sediments in lake catchments (Snowball and Thompson, 1990; Snowball et al., 1999). Commonly magnetic susceptibility reflects the concentration of magnetic minerals (e.g. Thompson and Oldfield, 1986). Cold climates without a stabilising vegetation cover and/or glaciers in the catchment cause high susceptibility due to increased erosion and deposition of minerogenic sediments (e.g. Stockhausen and Zolitschka, 1999; Nesje et al., 2000, 2001). Mineral magnetic measurements have been used, in combination with other physical sediment parameters, as an indicator of glacier activity (Sandgren and Risberg, 1990; Nesje et al., 1991, 1994, 2000, 2001; Karlén and Matthews, 1992; Matthews and Karlén, 1992; Snowball, 1993; Dahl and Nesje, 1994, 1996; Sohlenius, 1996; Matthews et al., 2000). Increased glacier activity in the lake catchment associated with increased erosion and input of clastic sediments increases the minerogenic content, also reflected in the LOI content. Periods of insignificant and reduced glacier activity in the lake catchment are characterised by low production of glacially derived clastic material. Increased magnetic susceptibility has therefore been related to the amount of allochthonous clastic material transport into lakes (e.g. Thompson et al., 1975). High magnetic susceptibility values may, however, also reflect increased input of minerogenic material into lakes associated with surface runoff during rainstorms, floods and mass movement events (snow avalanches and debris flows) from adjacent valley sides (Karlén and Matthews, 1992; Nesje et al., 1991, 1995, 2000, 2001; Dahl et al., 2003; Sletten et al., 2003).

3. Study area and previous research

Lake Danntjørn (8°99'E, 61°35'N), at an altitude of ~950 m, located in Sjødalen, eastern Jotunheimen, southern Norway, is a small mountain lake (maximum water depth of 4.1 m) located on top of a low-relief, NW/SE trending ridge just below the modern pine tree limit (~1000 m a.s.l.) in Sjødalen (Figs. 1 and 2A, B). The deepest part of the lake (the coring site) is located in the SE part of the lake. The lake has no major inlet streams and the outlet is in the SE end. Commonly the lake is ice free from May/June to September/October. Data on local hydrology, pH, summer lake water temperature and nutrient status are not available.

Mountain regions, such as Jotunheimen in central southern Norway, show large elevational differences over short horizontal distances. Jotunheimen is the highest mountain range in northern Europe and contains approximately 300 glaciers (Østrem et al., 1998). Most of Jotunheimen lies in the alpine zone, above the birch (Betula pubescens) altitudinal tree line at approximately 1000 m. Several mountains rise to more than 2000 m. Jotunheimen is located about 150 km from the western coast of southern Norway and the central part of Jotunheimen marks the main water divide in southern Norway. The climate in Jotunheimen is transitional between maritime western Norway and the
Fig. 1. Location map of southern Norway (left) and the Jotunheimen area (right). The contour interval is 100 m. The location of Danntjørn is indicated. Glaciers and lakes on the right-hand map are indicated by dark and light shading, respectively.
more continental eastern part of southern Norway. The meteorological station Sognfjell (1413 m a.s.l.) in western Jotunheimen had a AD1978–89 mean annual temperature of $3.1^\circ$C, a July mean of $5.7^\circ$C, a January mean of $10.7^\circ$C, a mean summer (1 May–30 September) temperature of $3.3^\circ$C and a mean winter (1 October–30 April) temperature of $10.7^\circ$C (Aune, 1993). The mean annual, winter, and summer precipitation were 860, 488 and 372 mm, respectively (Førland, 1993). The lowland region just east of Jotunheimen (Skjåk) is among the driest region in Norway with a mean annual precipitation of approximately 300 mm. The precipitation, however, increases by 8–10% per 100 m altitudinal rise (Haakensen, 1989; Dahl and Nesje, 1992; Laumann and Reeh, 1993), attaining >1000 mm over much of the high mountains. The lower altitudinal permafrost boundary is approximately 1460 m in eastern Jotunheimen and rises westward (Isaksen et al., 2002). The bedrock in Jotunheimen is predominately metamorphosed rocks of Caledonian age (charnockitic to anorthositic rocks). Along the valley bottom of Sjødalen, in which Lake Danntjørn is located, there is a belt of phyllite and sandstone of late Precambrian age (Sigmond et al., 1984).

Geographical and glaciological research in Jotunheimen has a history going back to the late 19th/early 20th centuries (e.g. Øyen, 1893, 1908). In western Jotunheimen, pioneering glacier mass-balance studies were carried out by Ahlmann (1922, 1940). In eastern Jotunheimen, glacial geomorphological processes were studied by Lewis (1960). Ice-cored moraines were studied by Østrem, 1964, 1965). Measurements and mapping of glacier-front variations were initiated by Hoel, Werenskiold and Liestøl (Hoel and Werenskiold, 1962) and annual glacier-front variations and glacier mass-balance data are published annually by Norges vassdrags- og Energidirektorat (NVE). The glacier mass balance record for Storbreen in central Jotunheimen, started in 1947, is the second longest and continuous record (after Storglaciären in Sweden) in the world (Andreassen and Østrem, 1999). In addition, Matthews (1974) used the Storbreen glacier foreland to contribute to the development of the statistical basis for licheno-metric dating of ‘Little Ice Age’ moraines. Recently, Matthews et al. (1997) reconstructed the Holocene colluvial history in Leirdalen, Matthews et al. (2000) used lake sediments from Bøvertunsvatnet and Dalsvatnet to reconstruct upstream glacier variations in western Jotunheimen, Barnett et al. (2001) reconstructed Holocene climatic change and tree-line response in Leirdalen, central Jotunheimen, and Sandvold et al. (2001) reconstructed the Holocene glacial and colluvial activity in Leirungsdalen.

4. Methodology

In summer 2000, two long, overlapping cores were retrieved from Lake Danntjørn from the deepest (~4.1 m depth) and flat-bottomed part of the lake (Fig. 2B). The long, 110-mm diameter cores were retrieved from a raft using a piston corer with a 110 mm diameter core tube constructed to obtain up to 6 m of sediments (Nesje, 1992). The upper core (core I) was 234 cm long, whereas the lower core (core II, top at 208 cm below lake bottom and basal part at 360 cm below lake bottom) was 152 cm long. The cores were brought unopened back to the laboratory where they were stored in a cold room until opening. After opening, the sediment layer closest to the tube wall was removed.

Fig. 2. (A) Map of the Danntjørn area. (B) Bathymetric map (contour interval 1 m) of Lake Danntjørn with the individual measurements (in metres) and the coring site indicated (x).
and the sediment surface was cleaned carefully. Lithofacies and sedimentological structures and textures were described before the cores were subject to magnetic susceptibility measurements and sampled contiguously for LOI at 0.5 cm intervals. In the summer of 2003, a 36-cm-long surface core was retrieved from a Zodiac rubber boat by a modified Kajak-gravity corer (Renberg, 1991; manufactured by HTH Teknik, Luleå, Sweden) from the same part of the lake as cores I and II. The Renberg gravity corer is designed to retrieve undisturbed cores from the water-sediment interface and downward in the upper part of the sediments.

The samples for LOI were dried overnight at 105 °C in ceramic crucibles before the dry weight was measured (normally 1–3 g). The water content was calculated in percent of the dry weight. In the furnace, the samples were subjected to gradually rising temperatures for half an hour and ignited at 550 °C for one hour. The crucibles were then put into a desiccator to cool for approximately half an hour and weighed at room temperature. The weight LOI was calculated in per cent of dry weight.

Magnetic susceptibility was carried out at 0.5 mm intervals on the cleaned core face of the two long sediment cores. The magnetic susceptibility measurements (10^-6 SI) of the two long Danntjørn cores were carried out by a Bartington MS2B sensor.

Core II was sampled contiguously (1-cm slices) between 311 and 355 cm (n = 45; corresponding to 232.5–277.5 cm in the composite record) for biogenic silica (BSi) content. BSi was extracted with 10% Na2CO3 and determined with spectrophotometer following Mortlock and Froelich (1989).

For radiocarbon dating, birch (Betula sp.) bark fragments were wet sieved (1-mm mesh) from the lake sediments. Accelerator mass spectrometry (AMS) radiocarbon dating (all 11 samples were bark fragments of birch) was carried out by the Poznan Radiocarbon Laboratory in Poland, following standard procedures for AMS radiocarbon dating. Calibration of the radiocarbon ages to calendar years BP (BP = AD 1950) was done using the calibration program CALIB 4.1.2 for atmospheric samples (Stuiver et al., 1998). When more than one possible intercept age, the median value was used. The age/depth model for Danntjørn lake record was based on linear interpolation between the intercept calibrated ages obtained from the different levels.

5. Results

5.1. Age/depth model

The depths in Table 2 refer to the depth below the sediment surface in the composite (core I and II combined, Fig. 5A–C) Lake Danntjørn sediment record, but the depths given in brackets refer to the original depths in core Danntjørn II (Fig. 5B). The date obtained

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Lab. ref.</th>
<th>¹⁴C age</th>
<th>Calendar year BP</th>
<th>Calendar year age range</th>
<th>Sedimentation rate (mm/year)</th>
<th>Time resolution (year/0.5 cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>57</td>
<td>Poz-932</td>
<td>2790±30</td>
<td>2915</td>
<td>2945–2850</td>
<td>0.44</td>
<td>12</td>
</tr>
<tr>
<td>91</td>
<td>Poz-930</td>
<td>3450±45</td>
<td>3695</td>
<td>3825–3640</td>
<td>0.39</td>
<td>13</td>
</tr>
<tr>
<td>119</td>
<td>Poz-931</td>
<td>3920±35</td>
<td>4410</td>
<td>4420–4295</td>
<td>0.31</td>
<td>16</td>
</tr>
<tr>
<td>152</td>
<td>Poz-936</td>
<td>2760±35*</td>
<td>2850</td>
<td>2920–2785</td>
<td>0.31</td>
<td>16</td>
</tr>
<tr>
<td>157.5 (235)</td>
<td>Poz-939</td>
<td>4980±40</td>
<td>5670</td>
<td>5745–5655</td>
<td>0.27</td>
<td>18</td>
</tr>
<tr>
<td>180</td>
<td>Poz-933</td>
<td>5720±40</td>
<td>6495</td>
<td>6550–6415</td>
<td>0.58</td>
<td>8</td>
</tr>
<tr>
<td>198.5 (276)</td>
<td>Poz-935</td>
<td>6060±40</td>
<td>6815</td>
<td>6970–6805</td>
<td>0.35</td>
<td>15</td>
</tr>
<tr>
<td>207</td>
<td>Poz-942</td>
<td>6200±40</td>
<td>7060</td>
<td>7210–7010</td>
<td>0.23</td>
<td>21</td>
</tr>
<tr>
<td>210.5 (288)</td>
<td>Poz-941</td>
<td>6260±40</td>
<td>7210</td>
<td>7250–7095</td>
<td>0.41</td>
<td>12</td>
</tr>
<tr>
<td>258.5 (336)</td>
<td>Poz-937</td>
<td>7555±45</td>
<td>8375</td>
<td>8390–8345</td>
<td>0.13</td>
<td>38</td>
</tr>
<tr>
<td>273.5 (351)</td>
<td>Poz-938</td>
<td>8550±50</td>
<td>9535</td>
<td>9545–9500</td>
<td>0.13</td>
<td>38</td>
</tr>
</tbody>
</table>

The depths refer to the composite (cores I and II combined) LOI record. Original depths in core II in brackets. For original depths in cores I and II, see also Fig. 5A, B. Dating marked* is not used in the age/depth model in Fig. 3. The ¹⁴C and calibrated age ranges are given with ± 1 sigma. All radiocarbon dates were obtained on birch (Betula) bark fragments. Calibration of the radiocarbon ages to calendar years BP (BP = AD 1950) was done using the calibration program CALIB 4.1.2 for atmospheric samples (Stuiver et al., 1998). When more than one possible intercept age, the median value was used. Sedimentation rates and time resolution between the radiocarbon-dated levels are indicated.
from 152 cm was apparently too young (the dated material may have been dragged downward during coring) and was therefore not used in the age/depth model based on linear interpolation (as recommended by R. Telford, pers. comm.) between the intercept calendar ages obtained from the dated levels (Fig. 3). Apparently from the age/depth curve, the topmost lake sediments were not sampled by the Nesje corer (core I), because the top sediments were extremely fluid and therefore difficult to recover. The age/depth curve was extrapolated from the three uppermost radiocarbon ages further upward in the core, indicating that approximately 60 cm (~1500 years) were lacking at the top. A surface core was retrieved in order to sample the time span lacking in Lake Danntjørn core I. Unfortunately, the entire time span could not be sampled (see below). The sedimentation rate (and time resolution) in the Lake Danntjørn sediment record was rather uniform (mean 0.37 mm/yr), corresponding to a mean time resolution of ~13 yr per 0.5 cm sample thickness in the upper 258 cm of the core (Table 2, Fig. 4). Below this level, however, the sedimentation rate was apparently significantly lower (0.13 mm/yr) corresponding to a time resolution in the order of 38 yr per 0.5 cm sample thickness.

5.2. Physical sediment parameters

The loss-on-ignition (LOI) records (contiguous samples at 0.5-cm intervals) from Lake Danntjørn cores I and II are shown in Fig. 5A, B. The LOI records from the Danntjørn cores I and II show the same main features (a–k in Fig. 5A, B) over the overlapping 77 cm sequence. Cores I and II from Lake Danntjørn were therefore combined into a composite LOI record (Fig. 5C). The tie points (based on LOI, magnetic susceptibility and age/depth models based on radiocarbon dates from the two cores) between the two cores were at 137 cm in core I and at 213.5 cm in core II (the vertical lines in Fig. 5A, B). The 0.5-cm interval magnetic susceptibility record (10^-6 SI) from cores I and II were spliced at the same levels as the LOI records, and plotted together with the record of weight percent residue after ignition at 550°C on a depth scale (Fig. 6). Except for the basal part, from approximately 250 cm and downward, where the two curves are mainly in phase, there is no apparent relationship between the residue curve and the magnetic susceptibility curve. The dry weight, LOI and residue after ignition at 550°C in the 36-cm long surface Lake Danntjørn core are shown in Fig. 7.

5.3. Biogenic silica (BSi) concentration

The BSi concentration was measured contiguously every cm between 311 and 355 cm (45 samples) in the lower part of core Danntjørn II, corresponding to 233–278 cm in the composite Lake Danntjørn record (Fig. 8A). The BSi content shows a marked transition at 260 cm (Fig. 8, upper panel). Below this level, the BSi values are apparently relatively high (~120 mg/g), whereas above this level, BSi concentration apparently is considerably lower (~50 mg/g). Independent LOI analysis was carried on the same samples as subject to BSi analysis (1-cm intervals) and they are highly correlated (r = 0.93) with the primary LOI analysis over the same depth interval. Below 260 cm the BSi concentration is apparently higher than above. When flux-corrected, however, the BSi concentration is somewhat lower below 260 cm (before ~8500 calendar years BP) than above (Fig. 8, lower panel). The BSi data show a strong negative correlation (r = -0.81) between LOI and BSi above 260 cm (~8500 calendar years BP) in the composite record, indicating that the sediments are diluted by BSi (diatoms giving low magnetic susceptibility and high residue/low LOI) resulting in a negative correlation between LOI (organic matter) and BSi (Fig. 9A, B).

6. Discussion

Standardised (individual values subtracted from the mean value for entire record and divided by the standard deviation) residue values after ignition at 550°C in the surface core show similar features as a tree-ring width series (standardised according to the
Fig. 4. Sedimentation rate (mm/year) and time resolution (yr/0.5 cm; sampling interval is 0.5 cm) between the dated levels in the Danntjørn composite record (Table 2) on depth (A, B) and calendar years BP (C, D) scales.
same procedure) from the Trøndelag area (Thun, 2002; Nesje et al., in preparation) about 200 km north of Jotunheimen (Fig. 10). This indicates that variations in the residue after ignition (probably mainly BSi from diatoms, as demonstrated in the depth interval 233–260 cm in the Lake Danntjørn composite sediment record) at 550 °C (not the LOI) are mainly controlled by summer temperature variations.

Fig. 11 shows the residue of the composite Lake Danntjørn sediment record and the surface core together with the composite magnetic susceptibility record according to the age/depth model. Apparently,
the first organic sedimentation in the lake occurred \( \sim 9900 \) calendar years BP. In the basal part of the core, up to about 8150 calendar years BP, the residue and magnetic susceptibility curves fluctuate mainly in phase. In this part of the core, the material also consists of significant (also when flux-corrected) amounts of BSi.
indicating that the residue mainly consists of detrital, minerogenic material and diatoms. Two significant maxima/minima in the residue/LOI curves at 260 and 255 cm, the first of which also represented by a significant magnetic susceptibility peak, date to 8500 and 8300 calendar years BP, respectively. Subsequent to approximately 8150 calendar years BP, the residue and magnetic susceptibility curves are decoupled because the residue after ignition most probably mainly consists of BSi, as demonstrated up to 233 cm (~7760 calendar years BP). The highest residue values occurred at 7800–7200, 7000, 6700, 6200, 5400–5050, 4300, 2500, 1800, 1550, 900, 700, and <100 calendar years BP. The most significant <8000 calendar years BP residue minima are recorded at 7100, 6800, 6600, 6000–5700, 5000–4400, 4200, 3800, 3600, 3200, 3000, 2750, 2300, 1900, 1700, 1070, 740, 600, 380, and 80 calendar years BP.

Several climatic implications follow from the temporal pattern of LOI variations shown in Fig. 12A. The Danntjørn lake record indicates major millennial-scale and minor centennial-scale events. These may relate <1400–1600-year periodicities/quasi-periodicities identified in other climate archives in the North Atlantic region and other regions (Bond et al., 1997, 2001; Stuiver et al., 1997; Campbell et al., 1998; Bianchi and McCave, 1999; Chapman and Shackleton, 2000; O'Sullivan et al., 2002; Gupta et al., 2003; Risebrobakken et al., 2003; Sarnthein et al., 2003).

6.1. 10,000–8500 calendar years BP

Between 10,000 and ~8500 calendar years BP, when the LOI in Danntjørn mainly reflects lake productivity (residue mainly in phase with magnetic susceptibility), organic content was high, indicating high summer temperatures (Fig. 12A). This is also seen in reconstructed summer temperatures from altitudinal variations in the pine-tree limit in the Scandes Mountains (Dahl and Nesje, 1996 and references therein), chironomid-inferred mean July temperature reconstructed from the Holebudalen site, northern Setesdalen, southern
Norway (Brooks, 2003), SST (August) (diatom-inferred) in the eastern Norwegian Sea (Vøring Plateau; Jansen and Koç, 2000; Birks and Koc, 2002; Koç and Jansen, 2002; Andersen, 2003; Andersen et al., 2004), and generally little (except for a peak around 9500 calendar years BP) drift ice in the North Atlantic (Bond et al., 2001).
Oxygen isotope records from planktonic foraminifera in the eastern Norwegian Sea (Vøring Plateau; Risebrobakken et al., 2003) do, however, not show such an early Holocene thermal maximum (Fig. 12).

6.2. The 8200 calendar years BP event

The two significant LOI minima/residue and magnetic susceptibility maxima present at 260 and 255 cm in the composite stratigraphy (Figs. 5 and 6), corresponding to ages of ~8400 and 8300 calendar years BP, are probably the same widespread Northern Hemisphere event as recorded in the GRIP (Dansgaard et al., 1993) and GISP2 (Grootes et al., 1993) Greenland ice cores, in lacustrine and proglacial lake records (Karlen, 1976, 1988; Karlen et al., 1995; von Grafenstein et al., 1998; Nesje et al., 2000, 2001; Nesje and Dahl, 2001), marine records (Bond et al., 1997; Klitgaard-Kristensen et al., 1998) and in speleothemes (Baldini et al., 2002). This climate oscillation was in south Norway termed the ‘Finse Event’ after the type-site at Finse north of the Hardangerjøkulen ice cap (Dahl and Nesje, 1994, 1996). Barber et al. (1999a,b) and Clarke et al. (2003, 2004) suggested that this ‘8200 event’ was triggered by a catastrophic drainage episode from glacial lakes south of the Laurentide ice sheet. The slight offset in the timing of the ‘8.2 ka’/‘Finse event’ in Lake Danntjørn may indicate that the age model at the base of the core is slightly skewed (oldest 14C dates slightly too old).

The period subsequent to the Finse Event/‘8.2 ka event’ until about 7200 calendar years BP, exhibits the highest residue (high BSi concentration) values during the entire Holocene in Danntjørn, indicating high summer temperatures during that time interval. This is also observed in reconstructed summer temperatures from altitudinal variations in the pine-tree limit in the Scandes Mountains (Dahl and Nesje, 1996 and references therein), chironomid-inferred mean July temperature reconstructed from the Holebudalen site, northern Setesdalen (Brooks, 2003), southern Norway, August sea-surface temperatures (SST) (diatom-inferred) in the eastern Norwegian Sea (Vøring Plateau; Jansen and Koç, 2000; Birks and Koç, 2002; Koç and Jansen, 2002; Andersen, 2003; Andersen et al., 2004) (Fig. 12B–D). Oxygen isotope records from planktonic foraminifera in the eastern Norwegian Sea (Vøring Plateau; Risebrobakken et al., 2003) did not indicate high temperatures during this time span (Fig. 12E) and this period was characterised by significant drift ice in the North Atlantic (Bond et al., 2001; Fig. 12F).

6.3. 7200–5600 calendar years BP

The Danntjørn record indicates variable but generally falling temperatures between 7200 and 6000 calendar years BP. Summer temperatures reconstructed from altitudinal variations in the pine-tree limit in the Scandes Mountains, chironomid-inferred mean July temperature reconstructed from the Holebudalen site, northern Setesdalen, August SST (diatom-inferred) in the eastern Norwegian Sea (Vøring Plateau) all show an almost similar development (Fig. 12B–D). Oxygen isotope records from planktonic foraminifera in the eastern Norwegian Sea (Vøring Plateau) indicate no temperature trend during this time span (Fig. 12E). This period was characterised by variable, but generally increasing, drift ice in the North Atlantic (Fig. 12F).

An inferred significant summer temperature minimum around 6000 calendar years BP is also seen (with
small time offsets) in a summer temperature record reconstructed from altitudinal variations in the pine-tree limit in the Scandes Mountains, chironomid-inferred mean July temperature reconstructed from the Holebudalen site, northern Setesdalen, southern Norway, SSTs inferred from diatoms in the eastern Norwegian Sea and a significant drift ice peak in the North Atlantic (Fig. 12).

6.4. 5600–5000 calendar years BP

In Danntjørn, high residue values, indicating high summer temperatures, were recorded between 5600 and 5000 calendar years BP. This interval was also characterised by high summer temperatures reconstructed from altitudinal variations in the pine-tree limit in the Scandes Mountains, in chironomid-inferred mean July temperature (slightly delayed probably due to dating inaccuracy) reconstructed from the Holebudalen site, northern Setesdalen, southern Norway, August SSTs in the eastern Norwegian Sea (Vøring Plateau), and a significant drift ice minimum in the North Atlantic. The oxygen isotope records from planktonic foraminifera in the eastern Norwegian Sea, however, do not indicate high temperatures during this time span.

6.5. 5000–4400 calendar years BP

Low residue values are recorded in Danntjørn between 5000 and 4400 calendar years BP, indicating low summer temperatures. This interval was also characterised by a period of low summer temperatures reconstructed from altitudinal variations in the pine-tree limit in the Scandes Mountains. This was also recorded in chironomid-inferred mean July temperature (slightly delayed probably due to dating inaccuracy) reconstructed from the Holebudalen site, northern Setesdalen, southern Norway, SSTs (diatom-inferred) in the eastern Norwegian Sea and presence of drift ice in the North Atlantic. The oxygen isotope records from planktonic foraminifera in the eastern Norwegian Sea, however, do not indicate low temperatures during this time span, despite some oxygen isotope minima in a couple of levels.

6.6. 4400–3900 calendar years BP

In Danntjørn, high residue values were recorded between 4400 and 3900 calendar years BP (peak value at
(A) July solar insolation at 30 deg. N

(B) Solar activity (d C-14 residual)

(C) Atmospheric carbon dioxide concentration

(D) GRIP, atmospheric methane concentration

(E) GISP2, volcanic sulphate (11-yr running mean)
4300 calendar years BP). This interval was also characterised by a period of high summer temperatures reconstructed from altitudinal variations in the pine-tree limit in the Scandes Mountains, high mean July temperature (slightly delayed probably due to dating inaccuracy) reconstructed from chironomids at the Holebudalen site, northern Setesdalen, August SSTs in the eastern Norwegian Sea. The drift ice record from the North Atlantic shows, however, some drift ice, though decreasing, during this time interval. Again, the oxygen isotope records from planktonic foraminifera in the eastern Norwegian Sea do not indicate high temperatures during this time span.

6.7. 3900 calendar years BP to the present

After the inferred temperature drop at 3900 calendar years BP in the Danntjørn record, low residue values prevailed until ca 2700 calendar years BP, followed by high values ~2400 calendar years BP. A similar development was observed in the record of summer temperatures reconstructed from altitudinal variations in the pine-tree limit in the Scandes Mountains, chironomid-inferred mean July temperatures at the Holebudalen site, northern Setesdalen, southern Norway, August SSTs in the eastern Norwegian Sea. The drift ice record from the North Atlantic also indicates significant drift ice during this time interval. The oxygen isotope records from planktonic foraminifera in the eastern Norwegian Sea show a more similar development than in the lower part of the core.

A residue minimum in the Danntjørn record ~1700–1600 calendar years BP is also recorded in the Scandes temperature record, chironomid-inferred mean July temperatures (though slightly delayed) at the Holebudalen site, northern Setesdalen, in the oxygen isotope records from planktonic foraminifera in the eastern Norwegian Sea, and as a period of significant drift ice in the North Atlantic.

For the remaining part of the Holocene, the records compared in Fig. 12, show mainly a similar temperature development. The summer temperature records discussed above, except the oxygen isotope records from planktonic foraminifera from the Voring Plateau. The reason for this discrepancy may be that the planktonic foraminifera mainly reflect the isotopic composition of the water masses below the thermocline, thus mainly reflecting a mean annual temperature signal than ‘pure’ summer temperature.

7. Possible climate forcing factors

There is growing evidence of millennial-scale variability of Holocene climate, at periodicities of ~2500 and 950 years possibly caused by changes in solar flux (e.g. O’Brian et al., 1995) and ~1500 years possibly related to an internal oscillation of the climate system (e.g. Bond et al., 1997).

A comparison between possible climate forcing factors (Fig. 13) and the Holocene records in Fig. 12, suggests that the early Holocene thermal maximum observed in most records was caused by increased summer solar insolation to the northern hemisphere. The sub-orbital millennial to decennial climate variability observed in the Holocene climate reconstructions was most probably a combined effect of solar activity changes, periods of increased volcanic aerosols, and internal feedback mechanisms in the earth’s climate system. Internal modes of the climate system variability (e.g. the North Atlantic Oscillation; Hurrell et al., 2003) may also have been responsible for some of the observed climate variability. In addition to external climate forcing factors, abrupt, catastrophic lake drainage from ice-marginal lakes during the termination of the last ice age had significant, regional climate impact (Barber et al., 1999a,b). The best example of this is perhaps the ~8.2 ka event’ detected in many palaeoclimatic archives that resulted from sudden glacial lake drainage at the margin of the Laurentide ice sheet (e.g. Clarke et al., 2003, 2004).

Most palaeoclimatic studies that discuss potential climate forcing factors have simply made correlations by curve matching (in the time domain) or by finding spectral peaks that correspond to similar frequencies (in the frequency domain) of one or more potential forcing factors (e.g. Bond et al., 2001). By the use of realistic climate models (coupled ocean-atmosphere models, preferentially also incorporating hydrological and vegetation feedbacks), it should be possible to combine high-resolution palaeoclimatic reconstruction, like those shown in Fig. 8, with climate models in order to test
the complex interactions between the different forcing factors to help explaining the primary forcing factors behind the observed Holocene climate development.

8. Conclusions

(1) Holocene summer temperature variations have been inferred from several sediment parameters in a lacustrine sediment record obtained from Lake Danntjørn, a small mountain lake located close to the modern pine-tree limit in Jotunheimen, central southern Norway.

(2) In the lake sediments from the deglaciation and early Holocene (below ca 260 cm; ~8500 calendar years BP), variations in residue after ignition at 550 °C and magnetic susceptibility fluctuated mainly in phase (positively correlated). The BSi was also relatively high in this part of the core. Thus, in the early part of the Holocene, the variations in the residue were primarily driven by input of detrital, minerogenic sediments, BSi (diatom) and terrestrial/lacustrine biological productivity in the lake itself and in the lake surroundings.

(3) LOI analysis throughout Lake Danntjørn cores I and II revealed that the variations in residue after ignition of the sediments at 550 °C in the upper 260 cm (<8500 calendar years) were not associated with variations in magnetic susceptibility, but most probably diatom productivity (BSi), as demonstrated between 8500 and 7760 calendar years BP.

(4) From approximately 8500 calendar years BP up to the present, variations in the residue most likely primarily reflected variations in low-minerogenic (low magnetic susceptibility with little variations), diatom (BSi) productivity. The summer temperature variability has been inferred from variations in residue after ignition at 550 °C in the composite Lake Danntjørn lake sediment record, which has a mean time resolution of ~15 years per 0.5 cm sample thickness.

(5) The highest residue values (probably mainly reflect high diatom productivity due to high summer temperatures) occurred at 7800–7200, 7000, 6700, 6200, 5400–5050, 4300, 2500, 1800, 1550, 900, 700, and <100 calendar years BP. The most significant <8000 calendar years BP residue minima (probably mainly reflect low diatom productivity due to low summer temperatures) were recorded at 7100, 6800, 6600, 6000–5700, 5000–4400, 4200, 3800, 3600, 3200, 3000, 2750, 2300, 1900, 1700, 1070, 740, 600, 380, and 80 calendar years BP.

(6) The inferred summer temperature record from Jotunheimen exhibits many similarities with other terrestrial Holocene summer temperature reconstructions and temperature reconstructions based on surface ocean proxies, whereas oxygen-isotope records from planktonic foraminifera, which mainly live below the thermocline at a few hundred metres water depth, from the eastern Norwegian Sea (Risembakken et al., 2003) deviate significantly from the other proxies; especially, during the first-half of the Holocene.

(7) A comparison between the most likely climate forcing factors and the Holocene records presented in this paper, may suggest that the early Holocene thermal maximum observed in most records may have been caused by increased summer solar insolation to the Northern Hemisphere. However, abrupt and high-amplitude climate oscillations prior to ~8000 calendar years BP (including the 8.2 ka event) may have been triggered by abrupt, catastrophic lake drainage from ice-marginal lakes during the termination of the last ice age (Barber et al., 1999a; Clarke et al., 2003, 2004; Nesje et al., 2004). Sub-orbital, millennial to decennial climate variability observed in the Holocene climate reconstructions, on the other hand, may have been a combined effect of changes in solar activity, periods of increased volcanic aerosols, and internal feedback mechanisms in the Earth’s climate system.

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References


Fennoscandian summers from AD 500: temperature changes on short and long timescales. Climate Dynamics 7, 111–119.


Sør-Norge. Norges vassdrags- og energiverk. Vassdragsdirektora-
tet, 88, 27–44.


Stockhausen, H., Zolitschka, B., 1999. Environmental changes since 13,000 cal BP reflected in magnetic and sedimentological properties of sediments from Lake Holzmaar (Germany). Quaternary Science Reviews 18, 913–925.


Thun, T., 2002. Dendrochronological constructions of Norwegian conifer chronologies providing dating of historical material. Doctoral Philosophical Thesis, Department of Biology, Faculty of Natural Sciences and Technology, Norwegian University of Science and technology (NTNU), 336pp.


