Environmental responses to the 9.7 and 8.2 cold events at two ecotonal sites in the Dovre Mountains, Mid-Norway

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Key words

Early Holocene; Paleoclimatology; 9.7 and 8.2 cold events; Scandinavia; Lake sedimentology; Varves; Biomarkers; Vegetation dynamics; Ecotones

Abstract

We found strong signals of two cooling events around 9700 and 8200 cal yrs. BP in lakes Store Finnsjøen and Flåfattjønna at Dovre, mid-Norway. Analyses included pollen in both lakes, and C/N-ratio, biomarkers (e.g. alkanes and br-GDGTs), and XRF scanning in Finnsjøen. The positions of these lakes close to ecotones (upper forest-lines of birch and pine, respectively) reduced their resilience to cold events causing vegetation regression at both sites. The global 8.2 event reflects the collapse of the Laurentide Ice Sheet. The 9.7 event with impact restricted to Scandinavia and traced by pollen at Dovre only, reflects the drainage of the Baltic Ancylus Lake. More detailed analysis in Finnsjøen shows that the events also caused increased allochtonous input (K, Ca), increased sedimentation rate, and decreased sediment density and aquatic production. br-GDGT-based temperatures indicate gradual cooling through the early Holocene. In Finnsjøen, ca. 3100 maxima-minima couplets in sediment density along the analyzed sequence of ca. 3100 calibrated years show the presence of varves for the first time in Norway. Impact of the 9.7 and 8.2 events lasted ca. 60 and 370 years, respectively. Pine pollen percentages were halved and re-established in less than 60 years, indicating the reduction of pine pollen production and not vegetative growth during the 9.7 event. The local impact of the 8.2 event *sensu lato* (ca 8420 – 8050 cal yrs. BP) divides the event into a precursor, an erosional phase, and a recovery phase. At the onset of the erosional phase, summer temperatures increased.

Introduction

The early Holocene is characterized by a series of well-documented climate instabilities, i.e. cooling episodes, that are likely driven by a slow reorganization of the North Atlantic thermohaline circulation (e.g. Andersson et al., 2004; Berner et al., 2010; Wanner et al., 2011) in combination with a decrease in summer solar insolation (Renssen et al., 2007) and probably also periodic presence of perennial Arctic sea ice cover (Stranne et al., 2014). The most prominent early Holocene cooling episodes ca. 11.300 cal yrs. BP (PreBoreal Oscillation: PBO), 9700-9300 cal yrs. BP (Erdalen 2 event *s.l.*), and 8200 cal yrs. BP (Finse event) are included in the quasi-periodic Holocene "Bond-cycles" (Bond et al., 1997). These climatic cycles are thought to be related to perturbations in solar radiation and/or continental ice sheet dynamics (Bond et al., 2001; Obrochta et al., 2016). The three cold periods are clearly recorded in the marine stratigraphy of the North Atlantic and Nordic Seas (e.g. Koç and Jansen, 1992; Haflidason *et al.*, 1995; Björck et al., 1997; Andersen et al., 2004; Berner et al., 2008, 2010), and by glacial deposits showing glacial readvances in Scandinavia and the Alps (e.g. Nesje and Dahl., 2001; Dahl et al. 2002; Bakke et al., 2005; Nicolussi and Schlüchter, 2012; Gjerde et al., 2016; Moran et al., 2016).

The impact of these climatic events in Europe, and in particular, the impact on vegetation is less clear. Obviously, due to the remains of the decaying Weichselian Ice Sheet lingering, the records of the earliest climate oscillation are sparse in Scandinavia (Björck et al., 1997; Paus et al., 2015) compared to further south (Bos et al., 2007; Dormoy et al., 2009). In contrast, the influence on vegetation from the 8.2 event is more frequently recorded in mid- and northern Europe than southern parts of the continent (Ghilardi and O'Connell, 2013; Fiłoc et al., 2017). Nonetheless, well-established cases of this event have been identified in Spain (Davis and Stevenson, 2007), SE Europe (Budja, 2007), and as far east as Syria (van der Horn et al., 2015). Few studies document the 9.7-9.3 event, and those that do only show minor changes in vegetation (Wohlfarth et al., 2004; Whittington et al., 2015; Burjachs et al., 2016). In context of the distinct 9.7-9.3 signals recorded in marine sequences and glacial deposits, the lack of vegetation responses of similar strength and frequency in continental Europe is surprising as the underlying mechanism is thought to be the same for all events. A possible cause for the fragmented records could be low sample resolution at some sites (Whittington et al., 2015), but most probably, the lack of studies at ecotonal sites could explain the limited vegetation signal for this event. It is at the vegetation boundaries, the ecotones, that vegetation is less resilient to climate change (Smith, 1965; Fægri and Iversen, 1989), and here the strongest effects of cold events are signaled. Today, numerical treatments of large pollen-data sets find regional patterns of vegetation and climate change (e.g. Seppä et al., 2007; Seddon et al., 2015; Hjelle et al., 2018). However, the number of sites may not be crucial for elaborating detailed geographical patterns of these events. More important is the quality of sites studied including their ecotonal positions.

This study compares multi-proxy records from two sites at Dovre (Norway): Flåfattjønna (Paus, 2010) and Finnsjøen, where the pollen data are reported by Thoen (2016). The aim is to shed new light on the question whether the early Holocene "Bond"-events impacted climate and vegetation in northern Europe. The lakes were close to ecotones (Flåfattjønna: upper pine-forest line, Finnsjøen: upper birch-forest line) during the Early Holocene, and show two short-lasting vegetation fluctuations during this period. To investigate the causes of these climatic oscillations, we use AMS-dates of terrestrial macrofossils and principle component analysis (PCA) of the Finnsjøen and Flåfattjønna pollen data, combined with XRF-scanning, elemental (C/N) ratio, and biomarker (glycerol diakyl glycerol tetraethers (GDGT) and *n*-alkane) analyses.

2. Regional setting

The Dovre mountain ridge with Lake Store Finnsjøen is situated between the valleys Drivdalen and Vinstradalen in Oppdal, Trøndelag County (Norway) (Fig. 1). Lake Flåfattjønna lies 30 km to the east of the ridge, in Tynset, Hedmark County (Fig. 1 and Paus et al., 2006; Paus 2010). Features of the two lakes and their surroundings are listed in Table 1. In continental areas such as the study area, the birch-forest line roughly follows the 10 °C July isotherm (Odland, 1996). Both lakes lie in the low alpine zone characterized by lichen-dominated dwarf-shrub tundra. In Drivdalen, 2 km west of Finnsjøen (Fig. 1), birch-forests reach ca. 1100 m a.s.l., and pine-forests ca. 900 m a.s.l. The region including Finnsjøen is renowned for its well-developed and species-rich flora that includes plants with a so-called centric distribution in Scandinavia. More details regarding environmental features of these sites are included in Paus et al. (2006, 2015) and Paus (2010).

3. Material and methods

We refer to Paus et al. (2006) and Paus (2010) for details on material and methods of the Flåfattjønna study. The Finnsjøen material and methods are described below.

3.1. Sampling and lithostratigraphy

The Finnsjøen lake sediments were cored at maximum water depth (14.7 m) from the ice-covered lake surface during winter. A 110-mm piston corer (Nesje, 1992) modified by A. Paus and J. Kusior (Dept. of Earth Science, Univ. of Bergen) was applied, which allowed us to use 6-m tubes and start sampling maximum 5 meters below the sediment surface. In Paus et al. (2015), results from the core section 1980-2195 cm below water surface were reported. The more detailed Holocene results presented in this paper, are based on the core section 1865-2040 cm depth below water surface showing distinctly laminated gyttja with numerous macrofossil and/or silty layers (Fig. 2). The analyzed sediments were described (Table 2) according to the method by Troels-Smith (1955). Sediments were cored in one continuous sequence. Hiati or correlated overlaps are absent in the core. However, during the XRF logging that requires length of sections less than 180 cm, one core section was cut one cm too long. Hence, 1 cm (2015-2014 cm depth below water surface) was removed from one core section. This 1 cm gap is shown as hiatus in Fig. 2.

3.2. Geochemical core logging

The loss-on-ignition (LOI) was measured in levels at 1-10 cm intervals in the studied core section. The sub-samples were dried overnight at 105 °C, weighed and ignited at 550 °C for 1 h. LOI was calculated as percentages of dry weight.

To document the sediment structure in the minerogenic part of the core, sediments were X-ray photographed using a Philips X-ray 130 kV instrument. The X-ray imagery was processed on a negative film, and thereafter transferred into a positive format using a digital camera. The resulting photos are found in Figs. 2 and 7.

The non-destructive ITRAX µXRF element core scanner from Cox Analytical Systems was applied to analyze variations in geochemical properties along the core surface as well as the colour- and the X-ray imaging. The core was scanned using a molybdenum (Mo) tube with a downcore resolution of 200 µm. The voltage and current were set to 35 kV and 50 mA, respectively, with a counting time of 10 s for each analytical step. The elements selected to represent downcore lithological variations are potassium (K) and calcium (Ca). In addition to elemental and colour scans of the core surface, a density graph was extracted from X-ray images and plotted versus depth with the same resolution like elemental analyses. Because X-rays penetrate through the core, they represent density and illustrate the overall layering that characterizes the sediment record.

3.3 Radiocarbon dating, varves and age-depth modelling

Eleven samples of terrestrial plant remains from the Finnsjøen sediments between 1810 and 2055 cm depth below water surface were AMS radiocarbon-dated (Table 3). All dates were reported as calibrated years BP (cal yrs. BP; present = AD 1950) based on the InCal13 calibration curve (Reimer et al. 2013). We converted the dates to calendar ages using CALIB 7.10 (Stuiver et al., 2017). The age-depth modelling (Fig. 3a) was obtained with the CLAM 2.2 R package (Blaauw, 2010) which recognized two outliers. The small variations in sedimentation rates displayed in Fig. 3a most likely reflect dating inaccuracies. When estimating pollen accumulation rates (PAR) and plotting different sedimentary features versus age (Figs. 2, 4, 5, and 7), we used the linear sedimentation rate estimated by interpolating between the oldest and youngest dates.

When Figs. 2 and 7 were enlarged, a microscale pattern of the XRF density graph and the Ca and K curves appeared showing couplets of alternating maxima and minima. For the density graph, we counted ca. 3100 couplets in the analysed sediment sequence spanning ca. 3100 calibrated years, which indicates the presence of annual varves (see discussion in section 4.1). However, the density curve reflected a floating chronology with no fixed attachment points to the radiocarbon chronology (Fig. 3b). Only minor differences were noted between the two chronologies. We have chosen to use the radiocarbon chronology to date the onset of events and estimating the sedimentation rates. On the other hand, the varve chronology was used to estimate the duration of events.

3.4. Pollen

Material for pollen analysis was sampled at 0.5–15 cm intervals between 1865 and 2040 cm depth. The samples were treated with HF and acetolysed according to Fægri and Iversen (1989). We added *Lycopodium* tablets to the samples (1 cm^3) for estimates of concentration and pollen accumulation rates (PAR; Stockmarr, 1971). Identifications were based on Fægri and Iversen (1989), Moore et al. (1991), and Punt et al. (1976-1996) in combination with a reference collection of modern material at University of Bergen. *Betula nana* pollen was distinguished using the morphological criteria of Terasmäe (1951). The pollen diagrams (Figs. 4, 5, and 6) were drawn by the computer program CORE 2.0 (Kaland and Natvig, 1993). In the pollen percentage diagram (Fig. 6), the calculation basis (ΣP) comprised the terrestrial pollen taxa. For taxon X of aquatic plants (AQP) and spores, the calculation basis was $\Sigma P+X$. We used the computer program CANOCO 4.5 (ter Braak and Smilauer, 1997-

2002) for detecting and plotting ordination patterns in the terrestrial vegetation development. The analysed data set included results from Lake Store Finnsjøen merged (using the option in CORE 2.0) with pollen results from the same time interval (7600-10.700 cal yrs. BP) in sediments of Lake Flåfattjønna, ca. 30 km east of Lake Finnsjøen (Paus, 2010; Paus et al., 2006).

Palynological terrestrial richness (PR) was estimated by rarefaction analyses (program RAREPOLL, Birks and Line, 1992) using the minimum sum of terrestrial microfossils (= 565) as the statistical base ($E(T_{565})$). Intermediate levels of disturbance maximize richness by preventing both dominance and extinction of species (Grime, 1973). In accordance with this, the low estimated terrestrial PR $E(T_{565})$ for abundant tree pollen (AP) (Fig. 4) should indicate closed forests, whereas local PR maxima indicated periods when the vegetation was positioned close to and above the forest-line (e.g. Aario, 1940; Simonsen, 1980; Seppä, 1998; Grytnes, 2003). However, at Finnsjøen pine-pollen was not a local signal (see section. 5.2.). To estimate local changes in palynological richness at Finnsjøen, we also estimated palynological richness ($E(T_{102})$) by subtracting the dominant regional pine pollen from the statistical basis (Fig. 4).

3.5. Biochemical characterisation

C/N analyses: The lake sediments were freeze-dried for 48 hrs to remove all traces of water. The freeze-dried samples were kept in a desiccator with 12M HCl (48 hours) to remove any traces of carbonates present in the sediments (Hedges and Stern, 1984). Elemental C and N were measured on a Perkin Elemental analyzer (2400 series II CHNS/O) for specific lake samples together with certified standards (Jet Rock, Svalvard Rock). The reproducibility of elemental analysis was ±10%.

Lipids: Plant waxes were extracted from ca. 1-4 g of freeze-dried sediment from selected intervals (see Fig. 2) with a mixture of dichloromethane and methanol (ratio of 9:1 by volume) by an automated solvent extraction (Dionex ASE 300). The total lipid extracts were injected into an Agilent 6890 gas chromatograph with a HP5-MS column (30 m× 0.25 mm internal diameter × 0.25 μm film). The oven temperature was kept constant at 35 ˚C for 6 minutes, increased to 300 ˚C at 5 ˚C min^{-1} and then held for 20 minutes. The chromatograph was coupled with an Agilent 5973 mass spectrometer and operated at 70 eV to scan the full range of charged particles from m/z 50 to 600 amu. High purity standards (S-4066) from Chiron (Trondheim, Norway) and deuterated compounds from Sigma-Aldrich (Munich, Germany) were used for quantification. The total input of higher odd *n*-alkane concentrations (*n*-C_{27}, C_{29} and C_{31}) was used to calculate the input of terrestrial plant waxes derived from higher plants. In addition, the ratio P_{aq} (Ficken et al. 2000) was calculated to estimate the input of waxes derived from in-lake algal production

$$P_{aq} = \frac{C_{23} + C_{25}}{C_{23} + C_{25} + C_{29} + C_{31}}$$

where C_n refers to *n*-alkane carbon chain length.

Glycerol dialkyl glycerol tetraether (GDGT): The total lipid extract of 11 of the Finnsjøen samples was re-dissolved in hexane/*iso*-propanol (99:1, v/v) and filtered using 0.45 μm PTFE filters. The branched and isoprenoidal GDGT distribution was analysed by high performance liquid chromatography/atmospheric pressure chemical ionisation – mass spectrometry (HPLC/APCI-MS) using a ThermoFisher Scientific Accela Quantum Access triple quadrupole MS. Normal phase separation was achieved using the method of Hopmans et al. (2016) that consists of two ultra-high performance liquid chromatography silica columns in tandem. Injection volume was 15 µL from 300 µL. To increase the sensitivity and reproducibility, all analyses were performed using the selective ion monitoring mode (SIM) to detect specific ions (*m/z* 1302, 1300, 1298, 1296, 1294, 1292, 1050, 1048, 1046, 1036, 1034, 1032, 1022, 1020, 1018, 744, and 653).

The relative abundance of 6-methyl over 5-methyl br-GDGTs is expressed as the IR_{6me} ratio (De Jonge et al., 2013).

$$IR_{6me} = \frac{IIa' + IIb' + IIc' + IIIa' + IIIb' + IIIc'}{IIa + IIa' + IIb + IIb' + IIc + IIc' + IIIa + IIIa' + IIIb + IIIb' + IIIc + IIIc'}$$

In addition, the branched versus isoprenoidal tetraether (BIT) index (Hopmans et al., 2004) that reflects the relative abundance of the major bacterial br-GDGTs versus crenarchaeol, an iso-GDGT likely produced exclusively by *Thaumarchaeota* (Sinninghe Damsté et al., 2002), was also quantified

$$\mathrm{BIT} = \frac{(Ia + IIa + IIa' + IIIa + IIIa')}{(Ia + IIa + IIa' + IIIa + IIIa' + crenarchaeol)}$$

4. Results

4.1. Lithostratigraphy

The sedimentation rate appears approximately linear showing an average growth of 0.56 mm/ year (or 17.7 years/cm) in the studied time interval 10.700-7600 cal yrs. BP (Fig. 3a). The core section 1865-2040 cm below the water surface consists of distinctly laminated to sub-laminated gyttja as indicated by the high-resolution colour scan and the density graph plot. The lamina observed by eye (Figs. 2 and 7) are normally 0.5-0.6 mm thick and occur as greyish silty horizons or as darker layers of distinct concentrations (amount) of macrofossils. The variability in the density graph also shows the laminated structure of the core confirming that the lamination is not only preserved in the top layer of the core, but is the structure of the entire core. Down-core density is plotted versus the number of electrons penetrating the core section for every 200 µm. The lower the cps number is, the higher the density. And the higher sediment density is, the higher minerogenic content in the core. The shift from minerogenic sediments to an increasing amount of biogenic components is clearly shown at ca. 10350 cal yrs. BP. The shift at 9800 cal yrs. BP to lower sediment density (higher cps) reflects transfer to a period with increased biogenic production and content. These depositional conditions dominated by higher biogenic production, characterise the period studied in this core. It is punctuated by short periods of increased minerogenic content, composed of higher density and/or lower productivity, centred around 9650, 9340 and 8200 cal yrs. BP (Figs. 2 and 7). Similarly, the relative concentration of potassium (K) and calcium (Ca) reflects the lithological variability with similar amplitude as expressed in the density graph. These lithological variations around the postulated cool periods are consistent with colour imaging indicating distinct shifts in colour. Notably, the lamina appear coarser and thicker than the warm periods (Figs 2 and 7). Because K and Ca represent particles from the local bedrock, the variability measured reflects shifts in allochtonous contributions. The major increases of K and Ca around cooling periods is centred around 9650 and 8200 cal yrs. BP and illustrates the sensitivity of this parameter to local environmental changes.

The lithostratigraphy also shows microscale laminations superimposed on the laminations observed by eye. Both for sediment density and the elements K and Ca there are densely shifting values where maxima alternate with minima forming couplets (Figs. 2 and 7). We counted 3117 density couplets over the 3100 calibrated years spanned by the analysed Finnsjøen sediments. This strongly points to the deposition of annual varves in Finnsjøen, here reported for the first time in Norway. The density maxima (i.e. low cps) reflect increased allochtonous minerogenic input during the thawing in spring/early summer, whereas the density minima (i.e. high cps) represent autochtonous organic production during summers (consistent with lower C/N and terrestrial organic matter input albeit representing low-resolution measurements). Hence, the varve origin appears as a mixture of clastic and biogenic factors (Zolitschka et al., 2015).

The clear-cut changes in K and Ca concentrations at the lower boundary of the 9.7 and the 8.2 cooling events are interpreted to represent gaps of 1.1–1.5 cm of the lake record due to climate influenced erosion of the underlying laminated units. These estimates are based on the counting and thickness estimates of lamina compared with the age model (Fig. 3a). The gaps are calculated to represent a removal of maximum 12 years of sediments at the beginning of the 9.7 event and maximum 20 years of sediments at the beginning of the 8.2 event. Obviously, this reflects a source of error for establishing a reliable varve chronology.

4.2. Pollen results and statistical analysis

We identified 47 terrestrial taxa in 47 levels of the Finnsjøen sediment-section. The pollen sum of terrestrial taxa analysed per sample varied between 535 and 2066 (mean ΣP: 1035). Seven local pollen assemblage zones were defined by visual inspection (Figs. 2, 5 and 6, Table 4). In five PAZ (S-2 to S-6), pine dominates showing values of 40-90% ΣP. During this period of pine maximum, there are two distinct and short-lasting pine minima of 40-50% ΣP (PAZ S-3 and S-5). At the same time, *Betula, Juniperus* and algae show percentage maxima.

The merged data set from lakes Finnsjøen and Flåfattjønna was subjected to a DCA ordination that showed a gradient length of 1.70 SD. This suggested linear response curves. Hence, we chose PCA as an ordination technique. A preliminary PCA including the dominant and entirely regionally represented *Pinus* (see section 5.2.), condensed scatter plots, and axis 1 captured 62% of the variation in the data. To reduce the influence of pine and enhance the influence of local features, pine was included as passive in the PCA (Figs. 8 and 9). Palynological richness (PR) and LOI were added as environmental variables during the statistical assessment. In Fig. 8, the light-demanding pioneers are concentrated to the left with medium to low axis 2 values, along with PR. Deciduous trees (e.g. *Ulmus, Corylus*) and herbs (e.g. *Valeriana, Geranium*) on fertile soils are situated to the right and/or at high axis 2 values. *Pinus* and LOI occur to the extreme right.

4.3. Biochemical results

C/N ratio varies between 12-20 indicating inputs of lacustrine algal production and higher plants from the catchment typical of lacustrine environments (Das et al., 2008). Higher values coincide with the onset of the cold 9.7 and 8.2 events due to soil erosion and increased outwash of nutrients, before it declines with the intensification of colder temperatures. C/N gradually increases after the cold periods. This transition is most evident after the 8.2 event.

n-Alkane concentrations increase core upwards with inflection points coinciding with the 9.7 and 8.2 cooling events (Fig. 2), interpreted as terrestrial organic matter and aquatic input. A lower P_{aq} ratio suggests less algal productivity. The percentage of terrestrial organic matter (mainly plant waxes) declines sharply by nearly 20% after the onset of the cold events and recovers again after climate ameliorates and vegetation recovers in the catchment. The increase of terrestrial organic matter is larger during the post-9.7 warming than during the recovery after the 8.2 event.

Glycerol dialkyl glycerol tetraethers (GDGTs) are abundant biomarkers in most types of natural archives (Schouten et al., 2000; Schouten et al., 2013). Two types of main GDGTs are recognized; Archaea synthesize isoprenoidal (iso-GDGTs), whereas bacteria synthesize branched (br-GDGTs) compounds. In general, br-GDGTs are more abundant in terrestrial settings, whereas iso-GDGTs are more abundant in sediments from large lakes and marine environments (Hopmans et al., 2004). Iso-GDGTs can have between 0 and 8 cyclopentane rings, whereas crenarchaeol has four cyclopentane and one cyclohexane ring (De Rosa and

Gambacorta, 1988; Schouten et al., 2000; Sinninghe Damsté et al., 2002; Schouten et al., 2013). br-GDGTs can have between 0 and 2 cyclopentane rings and/or between 0 and 2 additional methyl groups at either the C5 and C6 position (De Jonge et al., 2013; Schouten et al., 2000, 2013; Sinninghe Damsté et al., 2000).

In mineral soils, peat, and lake sediments, the degree of methylation of br-GDGTs is correlated with mean annual air temperature (MAT), while the degree of cyclization of br-GDGTs and the relative abundance of 6-methyl br-GDGTs over 5-methyl br-GDGTs is correlated with the pH (e.g., Weijers et al., 2007; Loomis et al., 2012; De Jonge et al., 2014; Naafs et al., 2017; Russell et al., 2018).

In all 11 samples from Finnsjøen, taken between 2022 and 1870 cm below water depth, GDGTs were present in abundance. The GDGT distribution was dominated by bacterial br-GDGTs over archaeal iso-GDGTs. The dominant archaeal lipid was iso-GDGT-0 with a relative abundance over iso-GDGT-1, -2, and -3 above 90%. Crenarchaeol was present in low abundance and the BIT index, reflecting the ratio between br-GDGTs and crenarchaeol, varied between 0.98 and 1.00. The br-GDGTs were dominated by acyclic 5-methyl homologues Ia, IIa, and IIIa with the latter being the most abundant compound with a relative abundance over all br-GDGTs of ~ 30%. In all samples, 5-methyl br-GDGTs were the most abundant, but 6-methyl br-GDGTs were also present. The IR_{6me} ratio, reflecting the ratio between 6- and 5-methyl br-GDGTs, ranged from 0.3 to 0.4.

The dominance of br-GDGT over iso-GDGTs, the small size of the lake, as well as the broader biomarker distribution (see above) suggests that the majority of GDGTs are derived from the surrounding soils. As such we used the global mineral-soil based calibration to convert the br-GDGT distributions into temperature and pH estimates (De Jonge et al., 2014).

$$MAT_{mr}\ (^{O}C) = 7.17 + 17.1 \times \{Ia\} + 25.9 \times \{Ib\} + 3.44 \times \{Ic\}$$
$$-28.6 \times \{IIa\}\ (\mathrm{RMSE} = 0.46\,^{o}C\ )$$

$$pH = 7.15 + 1.59 \times CBT' \qquad (\text{RMSE} = 0.52)$$

$$CBT' = \log \frac{(Ic + IIa' + IIb' + IIc' + IIIa' + IIIb' + IIIc')}{(Ia + IIa + IIIa)}$$

The MAT_{mr}-based temperatures range between 3 and 6 ± 4.6 ºC. Temperatures gradually decrease along the core with highest temperatures recorded in the oldest samples. The pH was relatively constant around 6.5 and mimics the temperature decline with slightly higher pH values in the oldest samples (Fig. 2).

5. Discussion

5.1. Background climate

The low-resolution biomarker data based on *n*-alkanes and br-GDGT provides complementary information about the organic matter sources and how the changes were driven by climate fluctuations.

The br-GDGT based terrestrial temperatures (MAT_{mr}) from Finnsjøen (Fig. 2) provide a general context of background climate in the region on which the "Bond"-events are superimposed. We explicitly assume that 1) br-GDGTs in the mineral soils surrounding the lake are the main source of these compounds accumulating in the lake sediments, and 2) br-GDGT distribution is biased towards the warmer season. It is hard to confirm these assumptions, but given that we do not detect the novel hexamethylated GDGT only known from lacustrine production (Weber et al., 2015), the small size of the lake, abundant presence of higher plant waxes and elevated C/N values, it is likely that most of the organic matter in the lake sediments is not derived from *in situ* production in the lake. At present, the region experiences temperatures well below freezing during winter months with average temperatures in January around -11.5 ºC (Table 1). Temperatures are on average 7.5 ºC in July. It is not clear whether br-GDGTs in soils that experience <0 ºC temperatures during part of the year are predominantly produced during the warmer season (Peterse et al., 2009; Weijers et al., 2011; Deng et al., 2016), but as bacterial growth is temperature dependent, it is likely that production of br-GDGTs in top soils is dominated by production when temperatures are above freezing, before being washed into the lake. The reconstructed temperatures for the early Holocene between 3 and 6 ± 4.6 ºC are 5 to 8 ºC higher than present-day annual mean temperatures of -2.5 ºC, further supporting a bias in br-GDGT production to periods when temperatures are above freezing. For comparison, the average of mean monthly temperatures above zero is today estimated to 3.5 to 4 ºC at the altitude of Finnsjøen. The temperature evolution with ~2 ºC higher MAT_{mr} around 10.000 cal yrs. BP compared to 8000 cal yrs. BP, follows the local summer insolation pattern (Fig. 2), providing additional evidence that the record is biased towards the warm season. Thus, it does not represent the annual mean temperatures. Our data supports the hypothesis that the Holocene thermal maximum (HTM) in Scandinavia occurred during the early Holocene, and may have occurred earlier than the pine maximum in this region.

The MAT_{mr} calibration error of ± 4.6 ºC and sample resolution prevent the identification of Bond-events in the temperature record. However, MAT_{mr} does provide information about the regional background climate, which was a few degrees C warmer than at present. This is consistent with palaeobotanical records from southern Scandes Mountains (Kullman, 2013; Paus, 2013; Paus & Haugland, 2017).

5.2. Regional pine pollen

The period in focus includes the Early Holocene pine maximum that is distinctly displayed in pollen diagrams from alpine areas in South-Scandinavia (e.g. Bergman et al., 2005; Bjune, 2005; Gunnarsdottir, 1996; Velle et al., 2005, Segerström and Stedingk, 2003). During this pine maximum, numerous megafossils show that the pine-forests reached their maximum elevation in south-Scandinavia (Selsing, 1998; Kullman, 2013; Paus and Haugland, 2017). These pine forests perhaps never reached much higher than 1105-1110 m a.s.l. in the study area because no megafossils are found above this elevation (Paus, 2010; Paus et al., 2011). According to Paus & Haugland (2017), pine-forests did not reach more than ca. 250 m higher than present forests during the pine maximum. This would imply an early Holocene pine forest-line at ca. 1150 m a.s.l. at Dovre, which is ca. 100 m lower than the altitude of Finnsjøen. Pollen and macrofossil data from Råtåsjøen (1169 m a.s.l.), ca. 16 km SSE of Finnsjøen, supports this conclusion (Velle et al. 2005).

On the other hand, the pine sedaDNA (Paus et al., 2015), the extremely high pine PAR (45 10^3 grains cm^{-2} a^{-1}), and the high pine pollen percentages (90 % ΣP) in sediments of Finnsjøen (1260 m a.s.l.) could contradict this conclusion. We regard these evidences of local pine forests as doubtful based on the following arguments. Pine sedaDNA was only found in one core-interval (Paus et al., 2015) which could reflect long-distance transport of pine remains or single specimens of low-growing "Krumholz" pine that are currently found up to 1200 m a.s.l. at Dovre. The PAR values are ca. 40 times higher than the threshold for indicating local pine forests (Jensen et al.; 2007; Seppä and Hicks, 2006) and reflect extreme sediment focusing (Davis et al., 1984; see discussion in Paus et al., 2015). Lastly, lowland hillsides in Drivdalen (Fig. 1) where tree-birch and pine grow today, would have been important sources for long-distance pollen. Such pollen could be dominant when local pollen production was low. Moreover, it is macrofossils of *Betula pubescens* and not pine that are found in the Finnsjøen sediments (Table 3) indicating presence of birch-forests in adjacent areas. The pine derived pollen is nevertheless dominant in the lacustrine record. Perhaps the birch-forests were open and had low pollen-production. Hence, the representation of long-distance pine pollen was enhanced in the sedimentary record. It is well known that pine is represented by dominant long-distance transport in other pollen based studies from the Arctic-Alpine regions (Aario, 1940; Gajewski, 1995; Paus, 2000). With these interpretative constraints on the pine pollen signal, we reconstruct the following local vegetation and climate development for the Finnsjøen area.

5.3. General trends of local vegetation development

The PCA ordination (Figs. 8, 9) roughly displays gradients of vegetation density/soil thickness increasing towards the right (axis 1) and soil fertility increasing upwards (axis 2). At Finnsjøen, pollen from *Pinus* and the warmth-demanding *Corylus, Ulmus*, and *Quercus* shows the strong influence of long-distance pollen transport. Nevertheless, local successions can be distinguished. Species-diverse pioneer vegetation on shallow soils developed (PAZ S-1; lower left in Figs. 8, 9), and is followed by forests with *Betula pubescens, Populus tremula*, and (from 9300 cal yrs. BP) *Alnus incana* on more organic-rich soils (PAZ S-2 to S-6). Thereafter, tall-herb *Betula/Sorbus/Alnus* forests with e.g. *Valeriana, Geranium, Filipendula*, and *Urtica*, developed on the fertile soils in protected sites locally, whereas dwarf-shrub heaths expanded on wind-exposed ridges (PAZ S-7).

Within the same period (7600-10.700 cal yrs. BP), the local development at Flåfattjønna followed a similar pattern (Paus, 2010), but deviated chronologically in some successional stages. First, Flåfattjønna was deglaciated more than 800 years later than Finnsjøen (Paus et al., 2015), and therefore showed a lagged succession by a delay in leaching of soil minerals into the lake. Fig. 9 shows that pioneer plant communities on unweathered mineral-soils (PAZ F-2) developed ca. 10.700 cal yrs. BP at Flåfattjønna; a successional stage that was reached earlier at Finnsjøen (Paus et al., 2015). However, even if weathering and leaching of soils started just after local deglaciation, soil pH was still high at Finnsjøen in PAZ S-1 (Fig. 2). Second, even if pine-forests thrived at Flåfattjønna (1110 m a.s.l.) and did not at Finnsjøen (1260 m a.s.l.), the pollen record showed maximum pine values for a longer period at

Finnsjøen (ca 10.000 – 8000 cal yrs. BP) compared to Flåfattjønna (9700 – 8500 cal yrs. BP; Fig. 4). It is likely that Drivdalen (2 km west of Finnsjøen and 700 m a.s.l.; Fig. 1), where temperatures allowed pine to grow for a longer period than at higher elevations, was an important contributor to the regional pollen representation at Finnsjøen. This would result in a stronger and longer-lasting percentage signal at the high-altitude Finnsjøen with vegetation of lower local pollen production than at Flåfattjønna (cf. Aario, 1940; Ertl et al., 2012).

During the pine maximum, when Finnsjøen was situated close to the upper birch-forest ecotone, and Flåfattjønna was situated close the upper pine-forest ecotone, the two distinct episodes of reduced pine percentages occur around 9700 cal yrs. BP and 8400-8200 cal yrs. BP at both Finnsjøen and Flåfattjønna (Fig. 4).

5.4. The 9.7 cold event – Erdalen event 2

Around 9700-9600 cal yrs. BP in the Finnsjøen sediments, pine percentages, pine PAR, and LOI (Figs. 2, 4, 5 and 6) reach short-lasting minima, K and Ca element intensity and X-ray density (Fig. 2) reflect increased soil erosion and outwash resulting in increased lamina thickness, whereas *n*-alkanes show lowered input of both terrestrial organic matter and aquatic homologs (Fig. 2). The short-lasting C/N maximum is interpreted to reflect erosion and outwash of terrestrial organic matter, whereas declining C/N values show that colder conditions reduced terrestrial input more than the aquatic production (cf. alkanes of terrestrial organic matter *vs.* P_{aq}). The first part of the subsequent C/N rise reflects lower aquatic production, whereas the later rise shows a warming that increased the terrestrial input more than the aquatic production according to the *n*-alkane trends.

In PAZ S-3 of the Finnsjøen pollen diagram, constituting the three-level pine percentage minimum (Fig. 6), PAR values of *Betula*, *Juniperus*, and *Salix* show little change from the previous S-2 (Fig. 5). Hence, their S-3 percentage maxima reflect the reduction of pine entirely represented by regional/long-distance pollen (see section 5.2.). In addition, after removing regional pine from the calculation basis, palynological richness shows no distinct changes (Fig. 4). However, PCA with pine removed from the data set, shows that the local S-3 vegetation returned towards previous pioneer stages of S-1 (Fig. 9). Altogether, the stratigraphical trends indicate the 9.7 changes as a cold event that influenced lacustrine and terrestrial productivity. However, the GDGT temperature estimates show no distinct changes (see section 5.1).

The onset of the 9.7 cold event is signaled by both regional (i.e. decline in pine) and local (e.g. Ca and C/N increase) parameters. The floating varve chronology (see section 4.1) suggests that LOI decreased ca. 10 years later than pine. This delayed LOI decrease might reflect the time needed to erode top soil within the lake's catchment. The soil-independent algae (cf. *Pediastrum*) took advantage of nutrients washed out during the cold event. They flourished around the same time as regional pine abruptly increased (Figs 2, 5 and 6) both trends support the onset of climate warming during this period. Local vegetation regrowth and soil formation, shown by increasing LOI, lagged climate amelioration by 10-15 years according to the varve chronology. The upper boundary of the dark eroded layer occurs when LOI reached pre-9.7 values.

According to the floating varve chronology (Figs. 3b, 7), the impact of the 9.7 cold event lasted ca. 56-58 years. Increased soil erosion and outwash during the event seem to have increased varve thickness above the average sedimentation rate of 0.56 mm/yr to a rate of ca. 0.77 mm/yr. Accordingly, the influence on regional pine lasted for ca. 56-58 years before pine pollen production recovered. Missing sediments from this section could add maximum 12 years to the duration of the 9.7 impact (see section 4.1). Most probably, a period of about 60-70 years is too short for pine forests to recover totally after being decimated by very cold conditions (cf. Kullman, 1986, 2005). We think that the distinct pine oscillation reflects a multi-decadal cold period whereby pine survived, but experienced reduced pollen production. According to Dahl et al. (2002), the 9.7 glacial advance at Jostedalsbreen, ca. 150 km WSW of Finnsjøen reflects a cooling of at least 1°C. This would have reduced the pine pollen production by a similar magnitude as displayed in the Finnsjøen pollen diagram (cf. Hicks, 2006).

Notably, at Flåfattjønna, an erosion layer distinctly reflects the 9.7 event. This layer including pine seeds and needles shows that pine pollen percentages and LOI decrease after the decline in pine PAR (Paus, 2010). The pine percentage maximum at the pine PAR minimum (Fig. 4) reflects outwash of soils containing remains of local pine (pollen and macrofossils) when regional and local total pollen production was reduced (Paus, 2010). In addition, at Flåfattjønna, PCA indicates regression of local vegetation towards pioneer stages during the 9.7 event (Fig. 9).

According to the radiocarbon chronology (Fig. 3a), this cold event occurred ca. 9605-9675 cal yrs. BP interpolated (Figs. 2), but the varves suggest the duration to be around 56-58 years (Fig. 7). At Flåfattjønna, the cold event is predicted based on fewer ^{14}C dates (Fig. 4). This low-resolution and inaccurate chronology dates the event to ca. 9500-9700 cal yrs. BP (Paus, 2010).

5.5. The 9.3 cold event

At both Flåfattjønna and Finnsjøen, the post-9.7 warming initiated a vegetation closure reaching the Holocene maximum according to total PAR (Fig. 4). At Finnsjøen, the vegetation closure strongly reduced palynological richness. This warming also initiated the rapid establishment of broad pine-forest belts in mid-Scandinavia, reflecting the July mean Holocene maximum (Paus and Haugland, 2017). Shortly thereafter, during the first half in Finnsjøen PAZ S-4 (Fig. 6), a small-scale cooling parallel to the 9.7 event is detected around

9300 cal yrs. BP (at 1963 cm depth, see also table 2), showing a minimum in sediment density and decreasing pine, and a delayed increase in freshwater algae (*Pediastrum, Botryococcus;* Figs. 2, 6). Furthermore, local vegetation became more open shown by decreasing total PAR (Fig. 5), increasing light-demanding shrubs and dwarf-shrubs (Fig. 6), and an increase in long-distance pollen (cf. *Corylus*). We think these changes reflect a climate cooling, though its local effect was less extensive than the 9.7 impact at Finnsjøen. As for the 9.7 event, the delayed increase in freshwater algae could indicate the warming following the short-lasting cold event. The 9.3 cooling reflects the collapse of the Laurentide Ice Sheet (Yu et al., 2010; Gavin et al., 2011) with distinct impacts in Canada (Gavin et al., 2011), Greenland (Young et al. 2013), Iceland (Brynjolfson et al., 2015), and further south at the Iberian Peninsula (Burjachs et al., 2016; Iriarte-Chiapusso et al., 2016). The 9.3 event is also recorded in the Greenland ice-cores (Vinther et al., 2009; Rasmussen et al., 2014). Furthermore, signals of the 9.3 south of Svalbard are weak (Werner et al., 2016). This, in line with the scarcity of Fennoscandian 9.3 records in eastern Europe and its limited impact at Finnsjøen, indicate that the 9.3 cooling had its main influence in western and southern coastal Europe.

In Finnsjøen, the 9.7 event (Erdalen 2) and 9.3 event are recorded as two distinct separate events. We therefore emphasize that the 9.7 event, which seems to have a Fennoscandian origin, i.e. the drainage of the Baltic Ancylus Lake (Nesje et al., 2004), is not formally a "Bond" event, and must therefore not be confused with the 10.3 or 9.3 "Bond" events of North Atlantic Ocean origin (Bond et al., 1997, 2001).

5.6. The 8.2 cold event – Finse event

At Finnsjøen, the post-9.3 changes with slight increase in vegetation density (Fig. 5), favored the moisture-demanding *Alnus* to expand within the area. Thereafter, stable records of vegetation and other parameters indicate a period of stable climate until ca. 8420 cal yrs. BP at the PAZ S-4/S-5 transition (Fig. 6), where colder temperatures are signaled. Here, pine percentages and PAR values decrease, probably due to a decline in summer temperatures that decreased regional pollen production (Paus and Haugland, 2017). The slightly later increase in tree-birch and alder PAR values (Fig. 5) could in addition reflect the lowering of vegetation belts within the region and the descent of the more warmth-demanding pine-forest. The rise in *Alnus* PAR indicates the presence of more moist and fertile soils in the region, whereas macrofossils (Table 3) show the local presence of birch-forests. The increase in juniper (Figs. 5 and 6) reflects more open vegetation locally, and PCA (Fig. 9) shows the recurrence towards more pioneer vegetation. The decreasing LOI and sediment density in the last part of S-5 (Fig. 2) point to increased soil erosion and outwash.

Around the S-5/S-6 transition at ca. 8225 cal yrs. BP, pollen data indicate harsher conditions showing a maximum in palynological richness (Fig. 4) and representation of pioneers (*Saxifraga oppositifolia* type, *Empetrum*) in a short period with low total PAR between the earlier birch and alder decrease and the later pine rise in the pollen diagrams (Figs. 5 and 6). Apparently, the birch-forest ecotone was lowered which displaced the local area towards the tundra vegetation zone. This opening of local vegetation occurs at the same time when erosion of terrestrial organic matter intensified, and further supported by the short-lasting maximum in C/N-ratio (Fig. 2). Thereafter, the deposition of a dark minerogenic layer (Figs. 2 and 7, Table 2) is initiated showing minimum sediment density, K, Ca, and LOI. This reflects maximum erosion and outwash due to deteriorating climate conditions. At the same time, an increase in pine in early S-6 (Figs. 5 and 6) reflects expanding regional pine-forests and/or increasing pine-pollen production. Both alternatives reflect warmer conditions during summer. A warmer lake could also be indicated by decreasing C/N values due to increased algal growth. This apparently contrasting evidence of climate points to the 8.2 weakening of the Atlantic current (Daley et al., 2011; Holmes et al., 2016) that resulted in an increased continental climate and enhanced amplitude of seasonal temperatures and involved at least colder winters (Alley and Ágústdóttir, 2005). Hence, even if widespread local areas were exposed to maximum freezing/thawing and erosion due to a period of less snow and more wind during colder winters, summers became warm enough to allow regional pine to expand and/or increase its pollen production.

The bio- and litho-stratigraphy in PAZ S-5 and S-6 shows a two-step climate deterioration. The first of moderate impact is mainly signaled by the biostratigraphy at the PAZ S-4/S-5 transition, from ca. 8420 cal yrs. BP, whereas the second and strongest period is mainly reflected by geochemistry and a dark erosional layer that deposited from ca. 8225 cal yrs. BP close to the PAZ S-5/S-6 transition. The varve chronology suggests that this dark layer spans a period of ca. 38 years (Fig. 7). Probably, missing sediments could add maximum 20 years to this duration (see section 4.1.), which indicates that climate deterioration intensified ca. 8245 cal yrs. BP or a few years later. This coincides approximately with minimum $\delta^{18}O$-derived temperatures in the Greenland ice core (Fig. 2).

According to Rasmussen et al. (2014), the 8.2 event started ca. 8250 cal yrs. BP which is close to the estimated age of PAZ S-5/S-6 transition and the deposition of the erosion layer. It is likely these abrupt stratigraphical changes signal the sudden outburst of Lake Agassiz and the strong meltwater pulse into the North Atlantic that caused colder, drier and windier conditions globally (Alley et al., 1997; Alley and Áugústdóttir, 2005). Most probably, the moderate changes ca. 8420 cal yrs. BP reflects a longer-term cooling upon which the 8.2 event is superimposed (Rohling and Pälike, 2005).

A third phase, representing a recovery phase, from ca. 8175 to 8050 cal yrs. BP, shows increasing sediment density, K content, LOI, and total PAR (Figs. 2 and 5), which reflects stabilizing soils and re-development of vegetation cover. This long-lasting re-establishment phase of more than hundred years indicates that conditions gradually improved. The exposed position of Finnsjøen could partly explain the slow regrowth locally. On the other hand, the short-lasting algal-maximum ca. 8100 cal yrs. BP could indicate further warming that differentiates an older and still rather cold and unfavorable phase from a younger and warmer phase. In total, 370 years elapsed extending from ca. 8420 to 8050 cal yrs. BP, before the Finnsjøen vegetation totally recovered from the 8.2 impact.

At Flåfattjønna, the low-resolution chronology displays a 400 years local response (ca. 8550 to 8150 cal yrs. BP; Paus, 2010) to the 8.2 (*sensu lato*) impact appearing in a two-step moist-dry pattern. However, both steps were regarded as cold periods (Paus, 2010). Similar two-step patterns are a wide-spread phenomenon as reported by other researchers (e.g. Nesje and Dahl, 2001; Ojala et al., 2008; Rasmussen et al., 2008; Filoc et al. 2017). According to Rasmussen et al. (2014), the impact of the freshwater impulse into the North Atlantic lasted ca. 150 years (ca 8250 to 8090 cal yrs. BP), but terrestrial sites show longer-lasting responses of varying lengths (Filoc et al., 2017). Although a varying degree of dating precision must be considered, one must expect that the duration of vegetation responses to the same impact depends on both the geographical position (e.g. N-S, E-W, distance from coast) and distance from ecotones. Study sites at ecotonal positions, i.e. being less resilient to disturbance (such as the Finnsjøen and Flåfattjønna sites), seem to show long-lasting responses to the 8.2 event. Such sites probably were more vulnerable to the background climate variations such as the 8.2 precursor (Rohling and Pälike, 2005). Most probably, ecotonal sites also show a slow recovery after cold events.

5.7. The 9.7 and 8.2 events compared

Both at Flåfattjønna and Finnsjøen, the impact of the 8.2 event *sensu lato* (ca 370-400 years) lasted longer than the influence from the 9.7 cooling event (60 to < 200 years). At Flåfattjønna, the distinct PCA responses (Figs. 4 and 9a) show that the 8.2 event had a stronger impact on local vegetation than the 9.7 event. At Finnsjøen, with a better stratigraphic time resolution than Flåfattjønna, the strength of the two events could be reflected by increased sedimentation rate of their erosional layers. According to the varve chronology, the sedimentation rate of the 9.7 erosional layer is ~ 4.4 cm during 56-58 years (0.79-0.76 mm/yr), whereas the 8.2 erosional layer show a sedimentation rate of ~2.8 cm during 38 years (0.74 mm/yr). On the other hand, the 8.2 erosion occurred at a higher successional stage with denser vegetation on more mature soils (Fig. 9), i.e. when local vegetation was at a distance from ecotones and more resilient to disturbance. In spite of that, erosion during the 8.2 event shows similar values as during the 9.7 cold spell. Hence, the 8.2 cold event appears to have been stronger than the 9.7 cold event in the Finnsjøen area, and probably also over larger regions as signals of the 9.7 event are scarce.

To estimate patterns of impact, it would have been of interest to compare these results with studies from a wider area. To the best of our knowledge, other pollen records of the 9.7 event in Scandinavia are absent, whereas pollen signals of the 8.2 event are sparse in alpine south Norway.

At Topptjønna, 1.7 km south of Finnsjøen, the 8.2 event also shows a moist-dry two-step pattern, but the study was carried out with a much lower time resolution than at Finnsjøen (Paus et al., 2011). In Jotunheimen, ca. 110 km SW of Finnsjøen, the 8.2 event appears as a short reduction in pine pollen (Barnett et al. 2001; Gunnarsdottir, 1996). The Leirdalen site of Barnett et al. (2001) shows a strong pine oscillation from 90 % to 45 % and back to 90 %, lasting less than 80 years. Pine stomata show the presence of pine during the pine pollen minimum. Otherwise, the pollen percentages of *Betula, Salix* and herbs increased. This has been interpreted as a short-lasting cold event that affected pine pollen production but not the vegetative growth locally (Hicks, 2006). The similar pollen-stratigraphical patterns during the short-lasting Finnsjøen 9.7 event have been interpreted similarly (see section 5.4). During the 9.7 event, Finnsjøen was lying in open birch-forests well above the pine-forest line (see section 5.2.), whereas the Leirdalen site (920 m a.s.l.) was situated in closed pine-forests during the 8.2 event and far below the upper pine-forest ecotone. Even being situated at opposite sides of an ecotone, the border between pine-forest and birch-forest, the two sites showed similar pollen responses to the weaker 9.7 and the stronger 8.2 cooling, respectively. This demonstrates how vegetation response is determined by both the impact and the ecotonal position of the vegetation cover (cf. Smith, 1965; Fægri and Iversen, 1989)

6. Conclusions

- The sediments of Finnsjøen and Flåfattjønna show exceptionally strong stratigraphical signals of the 9.7 and 8.2 cooling events. The positions of these sites close to ecotones (vegetation borders) were decisive in reducing their resilience against climate fluctuations.
- At Finnsjøen and Flåfattjønna, the impact of the 8.2 event was stronger than the 9.7 event.
- During the abrupt 9.7 cooling event at Finnsjøen, pine pollen percentages became halved and re-established in less than 60 years indicating that pine pollen production was severely reduced due to lower summer temperatures.
- At Finnsjøen, the 8.2 event *sensu lato* (ca. 8420 – 8050 cal yrs. BP) can be divided into a precursor lasting ca. 195 years, an erosional phase lasting ca. 50 years, and a recovery phase lasting ca. 125 years. At the onset of the erosional phase, summer temperatures increased.
- In the Finnsjøen sediments, weak signals indicate a cold spell at 9300 cal yrs. BP. Both the 8.2 and 9.3 events reflect collapses of the Laurentide Ice Sheet and represent two of the global “Bond” events. The 9.7 event most probably reflects the drainage of the Baltic Ancylus Lake hat had restricted regional impact.

- The XRF sediment density graph documents annual varves throughout the studied Finnsjøen sediments.
- br-GDGT-based temperatures are biased towards warmer seasons and indicate gradual cooling throughout the Early Holocene, following local summer insolation.
- C/N ratios indicate input of lacustrine algal production and higher plant matter from the catchment area.
- Higher C/N values coincide with the onset of cold events and declines with its intensification; C/N increases again after the cold period.
- Input of terrestrial organic matter (plant waxes) decreases during cold conditions followed by its steady increase afterwards.

Acknowledgements

We thank Espen Paus, Erling Straalberg, and Ståle Samuelshaug for help during coring. The ITRAX µXRF core scanner system at the EARTHLAB, Department of Earth Science, University of Bergen, was used for elemental analyses, color and x-ray image scanning of the core. We also want to thank two anonymous reviewers for valuable comments which significantly improved the manuscript.

Funding sources

This work was financially supported by the Olaf Grolle-Olsen and Meltzer University Foundations

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Figure and table captions

Fig. 1: Maps of the Lake Finnsjøen and Lake Flåfattjønna areas. Numbers show altitudes in m a.s.l.

Fig. 2: Selected sediment features from the Finnsjøen core displayed along the linear age/depth model. From left: X-ray colour image, pollen-assemblage zones (PAZ), XRF scanning results of sediment density, K and Ca (cps: counts per second), loss-on-ignition (LOI), *Pinus* and *Pediastrum* percentages, C/N ratios, *n*-Alkanes (terrestrial organic matter and aquatic input), br-GDGT-based estimates of pH and mean annual temperatures (MATmr), mid-month summer solar insolation 60 °N (Berger, 1978), and temperature deviations from present in °C based on ^{18}O values from the Renland ice core, Greenland (Vinther et al., 2009), The 9.7, 9.3, and 8.2 cold events are shaded. The 8.2 event (*sensu lato*) is displayed by a tripartite development: the early precursor from ca. 8420 cal yrs. BP, the erosional phase from ca. 8225 cal yrs. BP, and the

recovery phase from ca. 8175 to ca. 8050 cal yrs. BP. Stippled red lines show the one cm thick sediment slice missing from the core (see section 3.1).

Fig. 3: a): Age-depth relationship for the Finnsjøen sediments. Grey area illustrates the 95% probability range. Two outliers marked with bold crosses, are recognized. The average linear sedimentation rate (white line) represents the preferred chronology (see section 3.3).

b): The floating varve chronology based on microscale patterns of the XRF sediment density graph and compared with the radiocarbon based age-depth chronology. The youngest part of the varve chronology is tentatively attached to the uppermost level 7600 cal yrs. BP dated by the radiocarbon-based age-depth model.

Fig. 4: Comparison of selected features from the Finnsjøen and Flåfattjønna merged data set. The 9.7 and 8.2 cold events are shaded. Radiocarbon-dated levels are marked in the Flåfattjønna age column. Shaded curves are 10x exaggerations of the scale.

Fig. 5: Pollen accumulation rates (PAR) for selected Finnsjøen taxa. Shaded curves are 10x exaggerations of the scale.

Fig. 6: Pollen percentage diagram from Finnsjøen. Calibrated dates are shown as mean probabilities (Stuiver et al., 2018). Shaded curves are 10x exaggerations of the scale.

Fig. 7: Detailed data of the 9.7 and 8.2 events at Finnsjøen. Figure displays scanning results (Xray colour image, sediment density, the elements K and Ca, loss-on-ignition (LOI), *Pinus,* and temperature deviation (^{o}C) from present based on ^{18}O values in Renland ice core, Greenland (Vinther et al., 2009). To the right, the enlarged sediment densities

during the 9.7 and 8.2 erosion layers show couplets of alternating maxima and minima values representing varves. Shading highlights the 9.7 and the 8.2 erosion layers.

Fig. 8: Plot of pollen taxa along the first two axes of the PCA of the merged pollen data set from Flåfattjønna and Finnsjøen. Merged data set includes 121 samples, 108 terrestrial taxa. Eigenvalues axis 1: 0.5020, axis 2: 0.1578, axis 3: 0.0896, axis 4: 0.0502. In the analysis, *Pinus* was treated as a passive taxon whereas loss-on-ignition (LOI) and palynological richness (PR) were included as environmental variables. See section 5.3 for ecological interpretations of the axes.

Fig. 9: PCA of spectra from Flåfattjønna (a) and Finnsjøen (b). Pollen assemblage zones (PAZ) follow Paus (2010) and Fig. 6. Levels in PAZ S-4 are not encircled. Figures show the general vegetation development and the 9.7 and 8.2 impacts on vegetation in a two-dimensional gradient space. See section 5.3 for ecological interpretations of the axes. The data from the two lakes are from the same time interval: 7600 - 10.700 cal yrs. BP.

Table 1: General features of the sites studied. Local temperatures are extrapolated from the nearest meteorological stations (DNMI, 2016) using a lapse rate of 0.6 ^{O}C change per 100 m.

Table 2: Description of the Finnsjøen sediment lithology.

Table 3: Results of seven AMS dates of plant macrofossils from Finnsjøen. Calibrated dates according to Stuiver et al. (2018) are shown with two standard deviations. When dating results appear as two or more intervals, the two extreme values define the interval displayed. Median probabilities are shown in brackets. Lab. reference

numbers of two outliers are marked with [A]: ETH-48538 [A] and TRa-4470 [A]. Dates previously published (Paus et al., 2015), are marked with an asterisk.

Table 4: Names, dates, and biostratigraphical features of the Finnsjøen local pollen assemblage zones (PAZ).

Fig.1

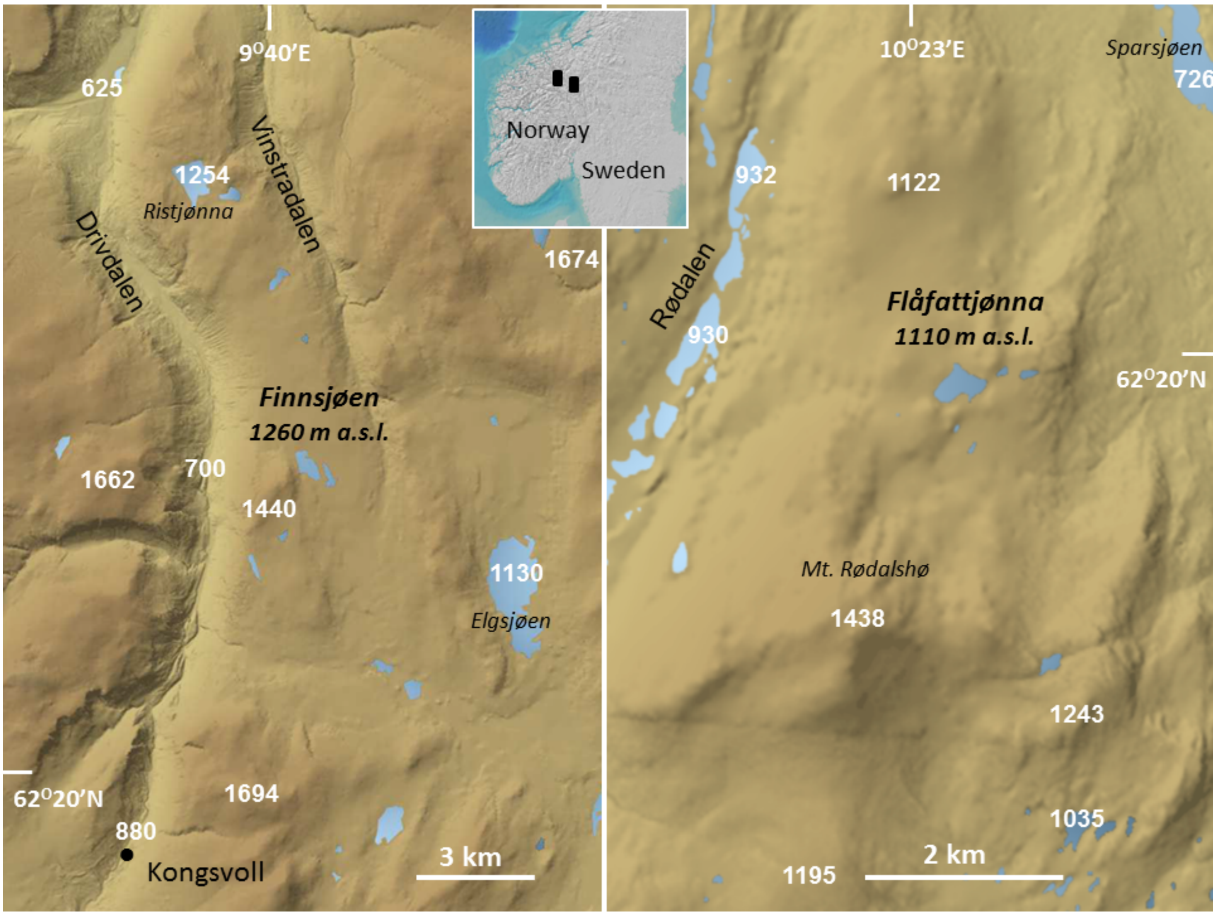


Fig.2

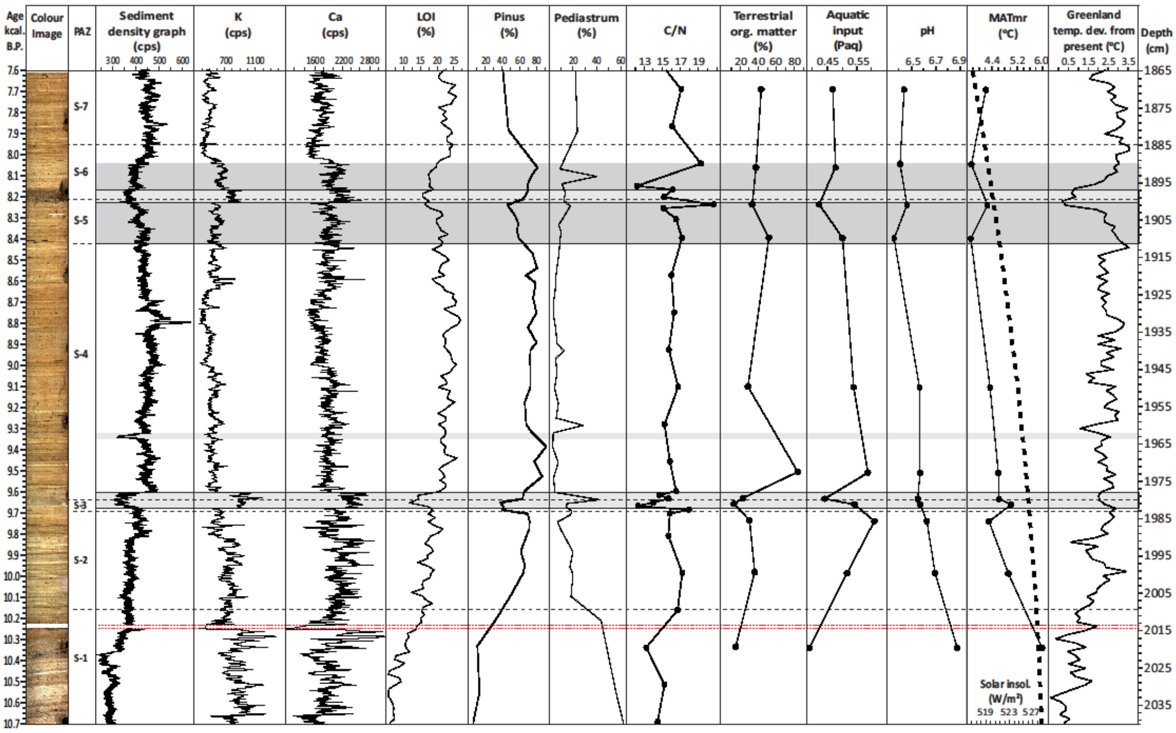

Age kcal. B.P.
Colour Image
PAZ
Sediment density graph (cps)
K (cps)
Ca (cps)
LOI (%)
Pinus (%)
Pediastrum (%)
C/N
Terrestrial org. matter (%)
Aquatic input (Paq)
pH
MATmr (°C)
Greenland temp. dev. from present (°C)
Depth (cm)
S-7
S-6
S-5
S-4
S-3
S-2
S-1
Solar insol. (W/m²)

Fig 3.a

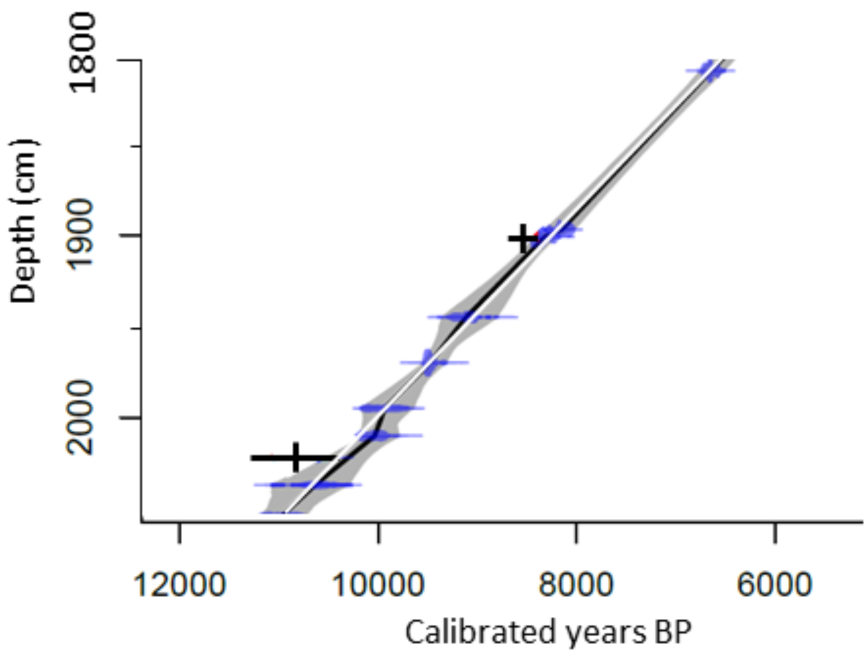
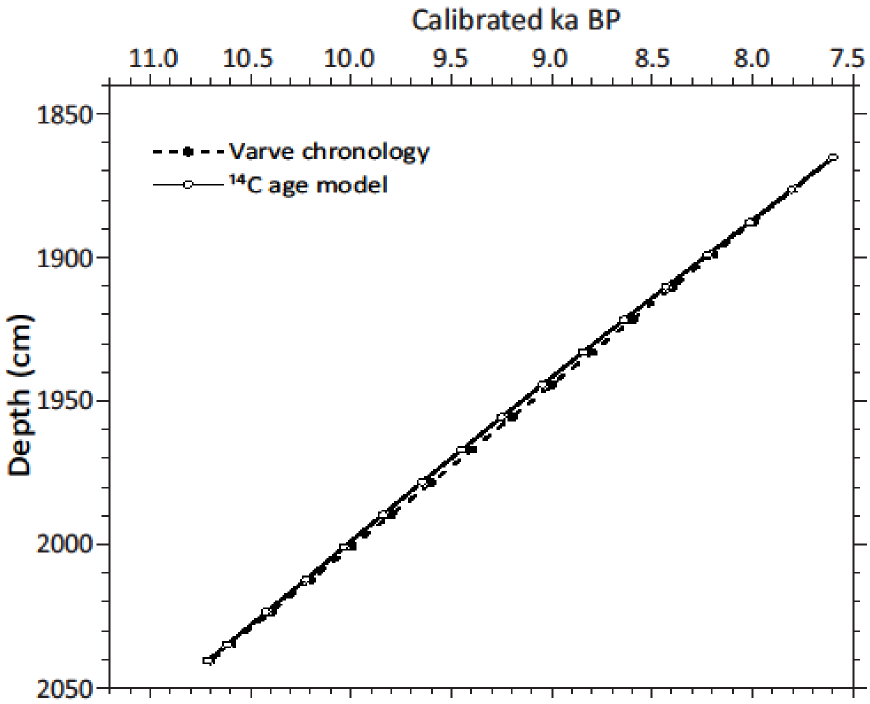

Depth (cm)
1800
1900
2000
12000
10000
8000
6000
Calibrated years BP

Fig. 3b



Depth (cm)	^{14}C Age model	Varve chronology
1865	7600	7600
1876	7800	7806
1888	8000	8015
1899	8200	8226
1910	8400	8433
1922	8600	8641
1933	8800	8847
1944	9000	9048
1956	9200	9250
1967	9400	9450
1978	9600	9648
1990	9800	9842
2001	10000	10038
2012	10200	10227
2024	10400	10424
2035	10600	10619
2040	10700	10717

Fig.4

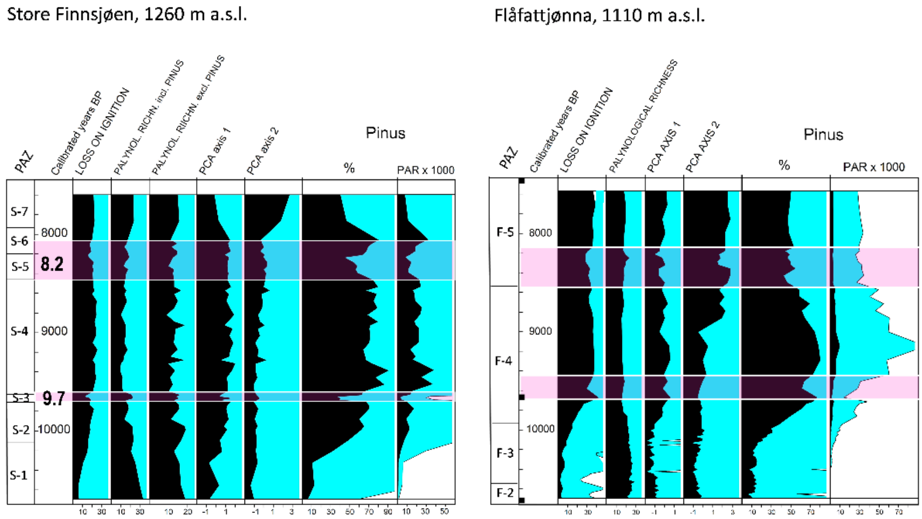

Store Finnsjøen, 1260 m a.s.l.
PAZ
Calibrated years BP
LOSS ON IGNITION
PALYNOL. RICHN. incl. PINUS
PALYNOL. RICHN. excl. PINUS
PCA axis 1
PCA axis 2
Pinus
%
PAR x 1000
S-7
S-6
S-5
S-4
S-3
S-2
S-1
8000
9000
10000
8.2
9.7
Flåfattjønna, 1110 m a.s.l.
PAZ
Calibrated years BP
LOSS ON IGNITION
PALYNOLOGICAL RICHNESS
PCA AXIS 1
PCA AXIS 2
Pinus
%
PAR x 1000
F-5
F-4
F-3
F-2
8000
9000
10000

Fig. 5

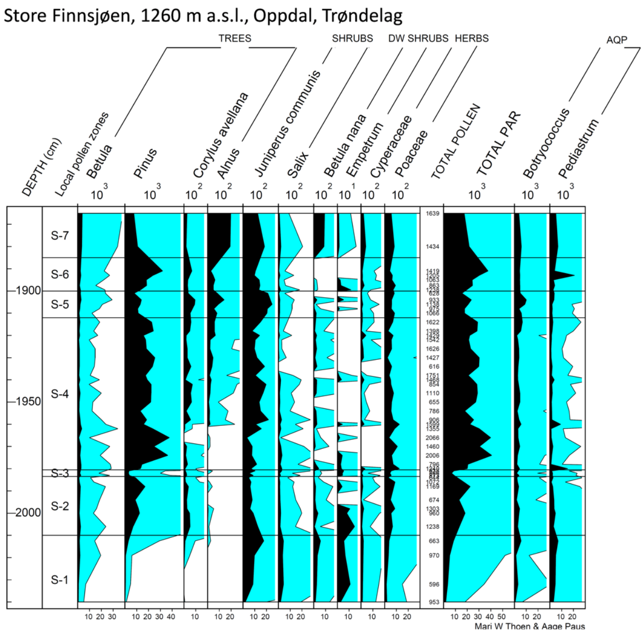

Store Finnsjøen, 1260 m a.s.l., Oppdal, Trøndelag
TREES
SHRUBS
DW SHRUBS
HERBS
AQP
DEPTH (cm)
Local pollen zones
Betula
Pinus
Corylus avellana
Alnus
Juniperus communis
Salix
Betula nana
Empetrum
Cyperaceae
Poaceae
TOTAL POLLEN
TOTAL PAR
Botryococcus
Pediastrum
S-7
S-6
S-5
S-4
S-3
S-2
S-1
1900
1950
2000
Mari W Thoen & Aage Paus

Fig. 6

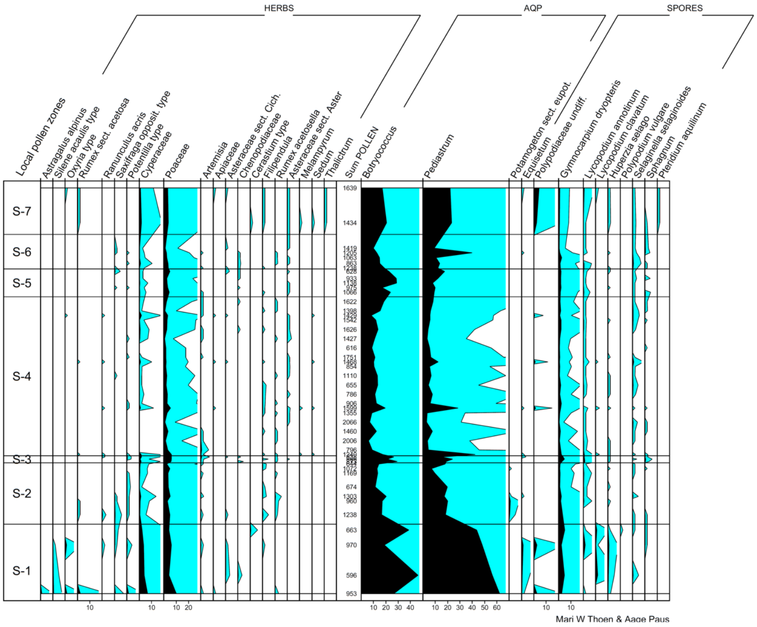

HERBS
AQP
SPORES
Local pollen zones
Astragalus alpinus
Silene acaulis type
Oxyria type
Rumex sect. acetosa
Ranunculus acris
Saxifraga opposit. type
Potentilla type
Cyperaceae
Poaceae
Artemisia
Apiaceae
Asteraceae sect. Cich.
Chenopodiaceae
Cerastium type
Filipendula
Rumex acetosella
Asteraceae sect. Aster
Melampyrum
Sedum
Thalictrum
Sum POLLEN
Botryococcus
Pediastrum
Potamogeton sect. eupot.
Equisetum
Polypodiaceae undiff.
Gymnocarpium dryopteris
Lycopodium annotinum
Lycopodium clavatum
Huperzia selago
Polypodium vulgare
Selaginella selaginoides
Sphagnum
Pteridium aquilinum
S-7
S-6
S-5
S-4
S-3
S-2
S-1
Mari W Thoen & Aage Paus

Fig. 7

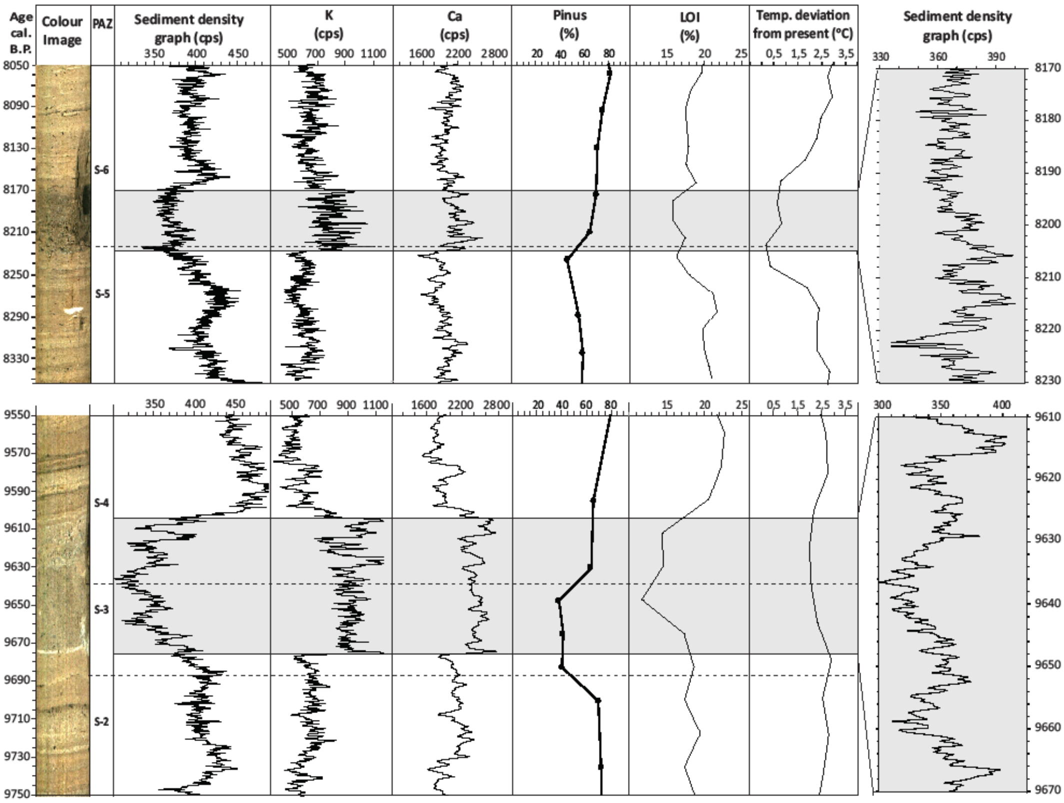

Age cal. B.P.
Colour Image
PAZ
Sediment density graph (cps)
K (cps)
Ca (cps)
Pinus (%)
LOI (%)
Temp. deviation from present (°C)
Sediment density graph (cps)
S-6
S-5
S-4
S-3
S-2

Fig.8

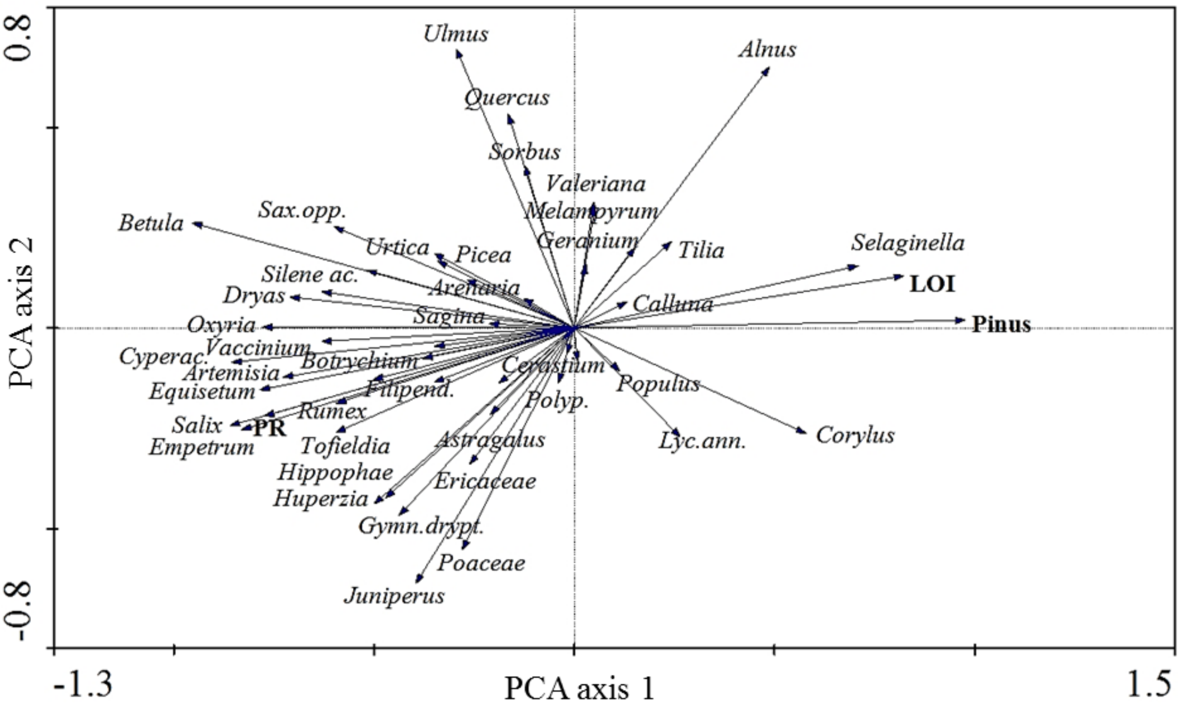

0.8
-0.8
PCA axis 2
-1.3
PCA axis 1
1.5
Ulmus
Alnus
Quercus
Sorbus
Valeriana
Melampyrum
Geranium
Tilia
Selaginella
LOI
Pinus
Betula
Sax.opp.
Urtica
Picea
Silene ac.
Dryas
Arenaria
Calluna
Sagina
Oxyria
Vaccinium
Cyperac.
Botrychium
Artemisia
Cerastium
Populus
Equisetum
Filipend.
Polyp.
Rumex
Salix
PR
Empetrum
Tofieldia
Astragalus
Lyc.ann.
Corylus
Hippophae
Ericaceae
Huperzia
Gymn.drypt.
Poaceae
Juniperus

Fig. 9a

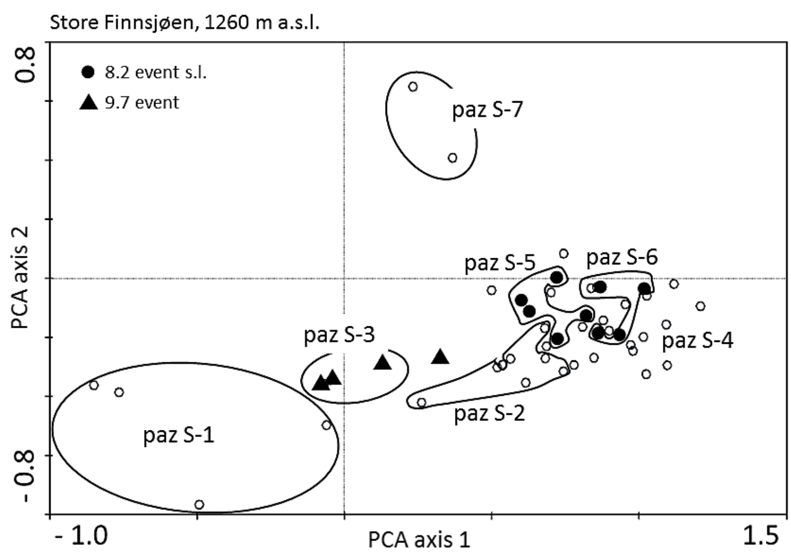

Store Finnsjøen, 1260 m a.s.l.
8.2 event s.l.
9.7 event
0.8
-0.8
PCA axis 2
paz S-7
paz S-5
paz S-6
paz S-3
paz S-4
paz S-1
paz S-2
- 1.0
PCA axis 1
1.5

Fig. 9b

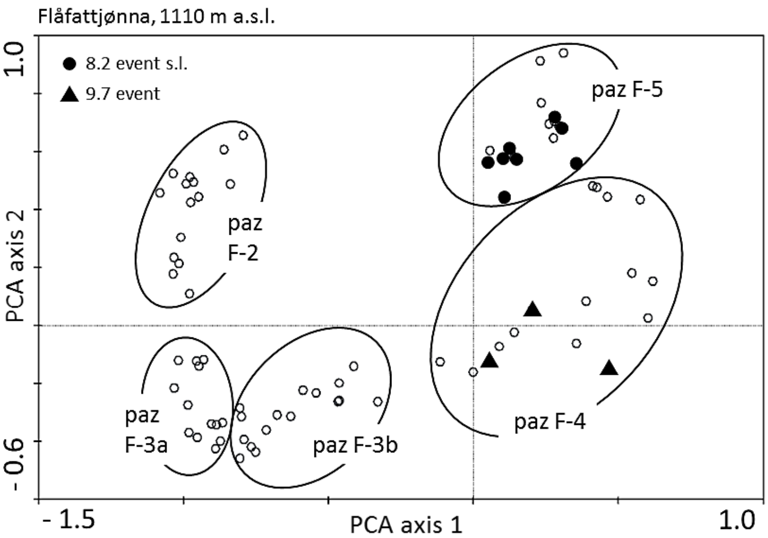


Table 1

Table 1

	Lake Finnsjøen (1260 m a.s.l.)	Lake Flåfattjønna (1110 m a.s.l.)
Geographical position	62°24’N, 9°41’E	62°20’N, 10°24’E
Coring point position UTM 32V NQ	0535133 E 6918753 N	0572506 E 6911883 N
Basin size Basin area	800m x 390m 23.7 ha	425m x 225m 6 ha
Maximum water depth	14.7 m	13 m
Catchment size incl. basin	69 ha	25 ha
No of inlets /outlets	0 / 1	0 / 1
Local bedrock	greenschists, slate, amphibolite	Phyllite, micashists
July mean	7.5 ^{O}C	9 ^{O}C
January mean	-11.5 ^{O}C	-13 ^{O}C
Annual mean	-2.5 ^{O}C	-1.5 ^{O}C
Annual precipitation	450 mm	500 mm
Local birch-forest line	1100 m a.s.l.	1030 m a.s.l.
Local pine-forest line	900 m a.s.l.	820 m a.s.l.

Table 2.

Table 2

Depth (cm)	**Description (Troels-Smith 1955)**	**Colour**	**Comments**
1865-1898	Ld^3 3, Dh 1, Ag +	Dark brown (nig 3÷)	Laminated gyttja. Less laminated in the upper part. Distinct laminations rich in macrofossils are found at 1893 and 1867 cm. One distinct silty lamina occurs at 1885 cm.
1898-1901	Ld^4 2, Dh 1, Ag 1	Dark brown (nig 3+)	Silty layer rich in macrofossils and without laminations. Shining from mineral particles.
1901-1978	Ld^3 4, Dh +, Tb +, Ag +	Dark brown (nig 3)	Laminated gyttja, brown - grey brown in silty laminations. Distinct macro-layers at 1911, 1922, and 1963 cm. Distinct silt layers at 1921, 1929, 1957, 1960, and 1971 cm. One sand lens at 1904 cm
1978-1983	Ld^3 3, Ag 1	Grey brown (nig 3÷)	Unstratified silty gyttja
1983-2021	Ld^3 4, Dh +, Tb +, Ag +	Dark brown (nig 3)	Laminated gyttja with macro remains
2021-2040	Ld^2 2, Dh 1, Ag 1, As +	Brown (nig. 2+)	Laminated clay/silt gyttja. Includes several dark (nig 3) macrofossil-layers less than 1 cm thick. Most distinct between 2021 and 2023 cm ($Ld^2$1, Tb1, Dh1, Ag+). Two mm thick and light (nig 1) clay layer at 2026 cm depth.

Table 3.

Table 4.

PAZ	Name	Age (cal. BP)	Pollen zone characteristics
S-7	*Alnus-Betula-Betula nana*	7580-7930	Pine declines to 45% ΣP and 10 10^3 grains cm^{-2} a^{-1}, respectively whereas *Alnus, Betula*, *Ulmus, Betula nana*, *Juniperus* and algae rise. Both palynological richness (PR) and LOI r
S-6	*Pinus-Betula*	7930-8270	Pine percentages rise earlier than pine PAR, both reaching max values (82% ΣP, 41 10^3 gra cm^{-2} a^{-1}, respectively) in mid S-6. *Alnus, Betula*, *Juniperus* and PR show distinct minima. I early S-6, LOI drops to 15% and rises to 24% in late S-6.
S-5	*Alnus-Betula-Juniperus*	8270-8520	Pine declines and reaches a minimum (50% ΣP, 3 10^3 grains cm^{-2} a^{-1}) in late S-5. *Alnus, Betula, Corylus*, and juniper show maxima. PAR values for all taxa rapidly drops in late S- LOI and PR show no changes from S-4.
S-4	*Pinus-Betula-Populus*	8520-9680	Pine strongly rises to its Holocene maximum (90% ΣP, 45 10^3 grains cm^{-2} a^{-1}) at 9.4 ka BP thereafter pine slightly decrease. At 9.4 ka BP, *Alnus* establishes. In S-4, LOI reaches 20-2 whereas *Betula, Salix*, and algae drop to moderate values. PR reaches its Holocene minimu
S-3	*Betula-Juniperus-Salix*	9680-9730	Pine abruptly decreases to 40% and 3 10^3 grains cm^{-2} a^{-1}, total PAR reaches a minimum of 10^3 grains cm^{-2} a^{-1}, and LOI drops to 14%. *Betula, Juniperus, Salix*, and algae show distinc maxima, but their PAR values show no changes. PR reaches a maximum of 26.
S-2	*Pinus-Corylus*	9730-10,070	Early S-2 shows marked increases in pine (65-70%), LOI (20%), and total PAR (48 10^3 gra cm^{-2} a^{-1}). Tree-birch, juniper, *Empetrum, Betula nana*, and algae decrease. PAR and PR decrease in the last half of S-2.
S-1	*Betula-Juniperus-Salix*	10,070 – 10,670	Sparse pine (< 35 % ΣP) and distinct representation of *Betula*, shrubs/dwarf-shrubs, and al characterize S-1. Total PAR (< 6 10^3 grains cm^{-2} a^{-1}) and LOI (< 15%) are low. Palynologi richness (PR) is high (24-33) and includes many light-demanding pioneer taxa.