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# Climate variability and lake ecosystem responses in western Scandinavia (Norway) during the last Millennium

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A B S T R A C T

This paper provides a high-resolution temperature reconstruction for the last Millennium from Lake Atnsjøen, SE Norway (61°52′31″N, 10°10′37″E). The sedimentary record reveals strong influence of the large-scale global climate patterns on the local climate in southern part of Eastern Norway. We reconstructed mean July air temperature using Chironomidae-based transfer function and fossil Chironomidae assemblages. The reconstruction was supported by a selection of climate-sensitive geochemical and paleoecological sedimentary proxies of terrestrial and aquatic origin, including Cladocera, pollen and macrofossils. Presented results revealed that summer temperatures were 1–2 °C warmer than the mean Millennial temperatures during the 11th, 13th, 15th and 20th centuries and 1–2 °C lower during the 12th, 14th, 17th–18th centuries. A persistent cold period, the Little Ice Age (LIA), occurred between 1550 and 1800 CE, was interrupted by a short warming at 1650 CE. The recognized regional climate fluctuations during the last Millennium affected the lake and its catchment, of which the strongest impact was caused by the LIA cooling. During the LIA the catchment vegetation was impacted by climate deterioration and the lake productivity reached its lowest level during the last Millennium. The current temperature reconstruction is in agreement with a previous continental scale temperature reconstruction for Europe. From obtained results it emerges that during the LIA the climate of western Scandinavia has been dictated by the atmospheric patterns originating from the North Atlantic.



## 1. Introduction

Climate variability and its impacts on the ecosystems can be reconstructed using natural archives, e.g. lake sediments that date back several millennia and provide multi-proxy long-term records (Battarbee et al., 2004; Elias and Mock, 2013). This is essential because a long temporal perspective on the magnitude and speed of environmental change is inevitable to discern between natural and anthropogenic forcing on the climate changes during the last centuries. A number of studies have been conducted to improve the knowledge of climate variability (Battarbee et al., 2004; Luterbacher et al., 2004; Mann, 2007; Nesje et al., 2000; Osborn and Briffa, 2006; Xoplaki et al., 2005) and several major global or hemispheric climatic episodes have been recognized during the last 1000 years (Mann et al., 2009; Osborn and Briffa, 2006; PAGES 2k Consortium, 2013; Wanner et al., 2008), among which the Medieval Climate Anomaly (MCA), the Little Ice Age (LIA), and the ongoing climate warming act as the most important.

Western Scandinavia is a key region to study paleoclimate due to its sensitivity to climate-forming processes in the North Atlantic shaping the climate of Europe (Helle, 2003). Of the large scale climate processes the North Atlantic Oscillation (NAO) is one of the main drivers of climate changes in Europe (Trouet et al., 2009; Wanner et al., 2008). It is speculated that the NAO is driven, at least partly, by solar forcing (Engels and van Geel, 2012; Shindell et al., 2001; van Geel et al., 1999). It has also been recognized that the external forcing related to atmospheric circulation patterns and solar activity cause long-term changes in aquatic ecosystem (Adrian et al., 2009; Dokulil et al., 2006; Luoto and Nevalainen, 2016; Ottersen et al., 2001). Though the global and regional climate trends are relatively well established, there is still a need for more detailed knowledge of the influence of large-scale atmospheric forcing on local climate dynamics at long temporal scales.

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The main aim of this research was to disentangle temperature variability and its impact on the functioning of a pristine lake in SE Norway with emphasis on climatic events related to changes in solar radiation and the NAO. The sediment core from Lake Atnsjøen located in the area with relatively low anthropogenic activity was analyzed. The reconstruction of climate changes and their impact on the lake environment during the last Millennium was performed using paleoecological and geochemical methods at ~30 yr resolution from a sediment core. The mean July air temperature was reconstructed using a Chironomidae-based calibration model and fossil Chironomidae assemblages, the lake and catchment responses were traced with indices based on subfossil Cladocera, pollen and macrofossils as well as total organic carbon (TOC).

The results were compared with a regional temperature reconstruction for the last two millennia (PAGES 2k Consortium, 2013), indices of solar activity (Reimer et al., 2013) and the NAO (Trouet et al., 2009) in order to examine the influence of large-scale external climate forcing on the local processes. The presented study aims at improving the understanding on local decadal climate dynamics and impacts during the last 1000 yr, their relationship with regional climate patterns, large-scale atmospheric and solar forcing.

### 1.1. Regional setting

Lake Atnsjøen is located in the eastern part of southern Norway, in the Atna Valley, to the east of the Rondane mountain chain (61°52′51 N, 10°09′55 E), 701 m a.s.l (Fig. 1). It is the largest lake in the River Atna watershed (area 4.8 km$^2$). The lake is deep and oligotrophic with a maximum depth of 80 m and a mean depth of 35.4 m. The river Atna drains into the lake in the western end, and leaves in the eastern. The retention time of the lake is about 6 months. The catchment of the lake Atnsjøen is 457 km$^2$ and the major part (85%) is located above the tree line (approximately 1000 m a.s.l.). The bedrock consists mainly of feldspar quartzite with locally large deposits of quaternary and fluvial materials (Halvorsen, 2004). The area has a continental, subarctic climate with average annual temperature and precipitation of 0.7 °C and 555 mm, respectively (Nordli and Grimenes, 2004). The mean July temperature is 13.4 °C (Fig. 2) based on the raw data provided by the Norwegian Meteorological Institute. A large part of the catchment lies within the Rondane National Park. Due to the remote location of the catchment, the Atna watershed and Lake Atnsjøen remain relatively unaffected by human activities.

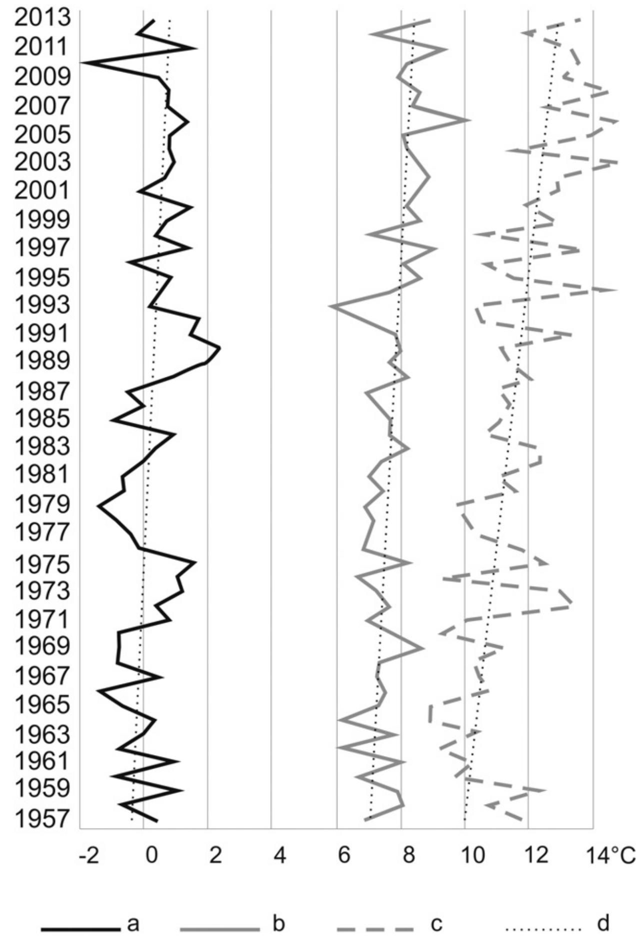


**Fig. 2.** Temperature data from Lake Atnsjøen for the period 1957–2012: a) annual, b) June–October period, c) mean July, d) mean value. The data are provided by the Norwegian Meteorological Institute and based on gridded temperature data (1 km$^2$) interpolated from surrounding weather stations. See Mohr (2008, 2009) for methods and more information.

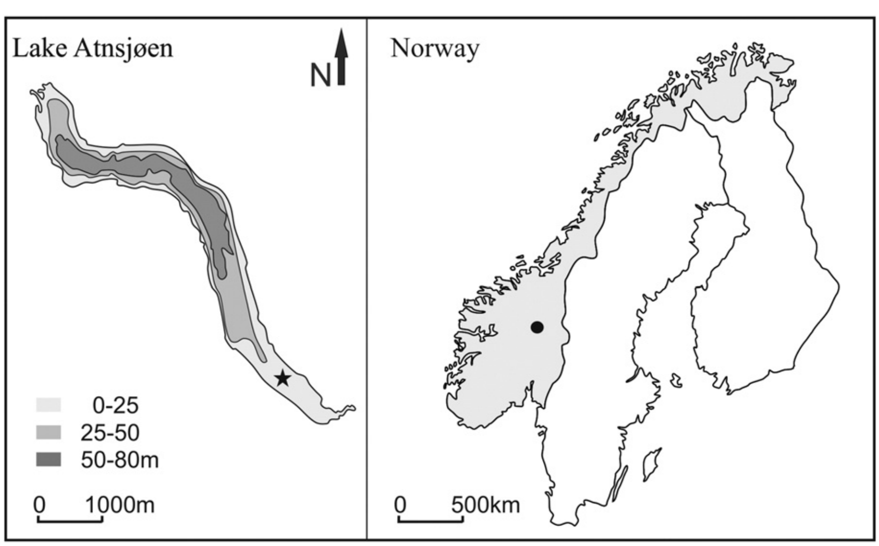


**Fig. 1.** Bathymetry of Lake Atnsjøen and its location in northern Europe, (a black dot). The bathymetric depth contours are grey shaded. The sediment core was taken in southern part of the Lake Atnsjøen and sampling location is indicated on the map with a black star.

## 2. Material and methods

A 34-cm sediment core from Lake Atnsjøen was taken in the south-eastern part of the lake, from water depth of 20 m using a KC-Denmark Kajak-type gravity corer (Fig. 1). The core chronology is based on $^{210}Pb$ distribution in the uppermost sediments, $^{14}C$ dating of terrestrial macrofossils and chronostratigraphic indicators inferred from pollen analysis. Multi-proxy analyses were performed with 1-cm resolution, which in average corresponds to 30 yrs. One cubic centimetre of fresh sediment was prepared in the laboratory according to standard procedures for each of the methods depicted below.

### 2.1. Chironomidae analysis

Subsamples for fossil chironomid analysis were prepared applying standard methods (Brooks et al., 2007). The wet sediment was gently sieved through a 100-μm mesh and the residue was examined using a Bogorov counting chamber under a stereomicroscope (32–40× magnification). Larval head capsules were extracted with fine forceps and mounted permanently with Euparal on microscope slides. Faunal identification was performed under a light microscope at 400× magnification. The minimum chironomid head capsule number per sample was set to 50. Identification of the chironomids was based on Brooks et al. (2007). Constrained unweighted pair-group average (UPGMA) cluster analysis was used to group samples into local chironomid zones. Bray-Curtis similarity index was used as a measure and the number of significant zones was assessed using the broken stick method.

### 2.2. Pollen and macrofossil analysis

The pollen and macrofossil analyses allowed us to deduce past vegetation changes in the catchment and in the lake. Pollen accumulation rate index (PAR) was used as an indicator of past abundances for trees (AP) and non-arboreal plants (NAP) separately (Theuerkauf et al., 2013), and interpreted as a climate change indicator. The AP index reflects regional and local vegetation changes while NAP represent composition of the local plant assemblages (Berglund and Ralska-Jasiewiczowa, 1986).

Samples for pollen analysis were prepared according to standard procedure (Berglund and Ralska-Jasiewiczowa, 1986). All sporomorphs were identified and counted until 500 pollen grains of trees and shrubs were reached. Pollen grains of trees were counted as arboreal pollen (AP) and all herbaceous species as non- arboreal pollen (NAP). The sum of AP and NAP was considered as 100%.

Samples for macrofossil analysis were prepared according to Birks (2007). Identification of fossil remains was based on the literature (Birks, 2007; Katz et al., 1977; Velichkevich and Zastawniak, 2008) and a reference collection. All macrofossil counts were standardized as numbers of fossils per 50 $cm^3$.

### 2.3. Cladocera analysis

The lake ecosystem response to climate changes was examined by the selected indices resulting from fossil Cladocera analysis (total Cladocera flux, bosminid sex ratio) (Fig. 6). Bosminid sex ratio was preferred over chydorid (Chydoridae) sex ratio because family Bosminidae dominated the core samples comprising 80–90% of all species. The productivity changes of the lake were traced using the total Cladocera flux, for which higher values indicate favourable conditions for organic matter to accumulate in the lake (Manca et al., 2007; Meyers and Teranes, 2001).

The fossil Cladocera and ephippium analyses were performed in the laboratory according to the standard procedure described in Frey (1986). Microscope slides were prepared from 0.1 ml of each sample and examined with a microscope (Olympus BX41). For each sample 2–6 slides were scanned and all skeletal elements: head shields, shells, postabdomens were counted. Identification of cladoceran remains was based on the identification key by Szeroczyńska and Sarmaja-Korjonen (2007). Total Cladocera flux calculations used absolute number of Cladocera individuals and number of years in the sample. For the ephippium analysis (to gain sex ratio values), bosminid (Bosminidae) carapaces and ephippia were identified and enumerated. The ratio of ephippia to carapaces was used as a bosminid sex ratio to indicate the relative significance of sexual reproduction. Fossil cladoceran ephippia are indicators for environmental stress, for example deteriorated climatic conditions, because sexual reproduction is induced by environmental stimuli, including lower temperatures (Kultti et al., 2011; Nevalainen and Luoto, 2013).

### 2.4. Total organic carbon

Sedimentary TOC was used as a lake biological productivity indicator (Meyers and Teranes, 2001). The coincidence between climate amelioration and the enhancement of biological productivity reflected by TOC in lakes has been reported (Anderson et al., 2008; Ramrath et al., 1999). TOC content in the sediment was analyzed from the dried and homogenized samples using an elemental CNS analyzer Vario Max CNS (Elementar). Prior to the measurements the samples were acidified with 1 M HCl to remove carbonates. The TOC flux [$mgC/cm^2/yr$] was calculated as TOC flux = TOC [mg/g] SAR [$g/cm^2/yr$].

### 2.5. Temperature reconstruction

To quantify natural climate variability, the mean July air temperatures using the Finnish Chironomid-based transfer function was reconstructed (Luoto et al., 2009). Temperature is clearly the most significant environmental determinant of chironomids in the training set with species-environment correlation of 0.90. In addition, conductivity, hypolimnetic oxygen and water depth have marginal influence on the assemblage composition. The model uses weighted-averaging partial least squares (WA-PLS) technique and has a cross-validated coefficient of determination of 0.78, a root mean error of prediction of 0.72 °C and a maximum bias of 0.79 °C. The model consists of 77 lakes spanning from boreal lakes to mountain birch woodland and tundra lakes. The mean July air temperature ranges from 17.0 to 11.3 °C (Luoto, 2009). Sample-specific errors (eSEP) were estimated using bootstrapping cross-validation (999 iterations). The suitability of the model to the fossil data was tested applying the modern analogue technique (MAT) to see whether the samples have good modern analogues ("poor" modern analogue > 5% chord distance). In addition, principal component analysis (PCA) axis 1 and 2 scores of the fossil chironomid assemblages were tested against the reconstructed temperature values using Pearson correlation and the level of statistical significance. Correlation between the reconstructed values and PCA axis 1 scores indicate that the community has been responding to the reconstructed variable, whereas temperature correlation with a secondary axis could suggest community response to other variables. The PCA was run with square-root transformed (relative abundance) assemblage data. To test whether there is significant offset in the reconstructed value for the surface sample, we compared it against the instrumentally measured mean July temperature (period 1957–2013). When the difference between the reconstructed and measured value remains below the eSEP of the surface sample, the reconstruction can be considered successful.

The results were compared to continental-scale temperature variability (PAGES 2k Consortium, 2013) and to other main paleoclimate reconstructions from Scandinavia and the north Atlantic region (Luoto, 2013; Mann et al., 2009; Osborn and Briffa, 2006). In order to reveal the mechanism behind the climate fluctuations in studied region the reconstruction was related to indices of solar forcing (Reimer et al., 2013) and the NAO (Trouet et al., 2009). The modern reconstructed temperatures from Lake Atnsjøen were compared with contemporary meterological data. These were gridded temperature data (1 $km^2$)

interpolated from surrounding weather stations. The data were provided by the Norwegian Meteorological Institute (Mohr, 2008; Mohr, 2009)

## 3. Results

### 3.1. Chronology

The age-depth model was based primarily on $^{210}$Pb dating results. The unsupported $^{210}$Pb distribution with mass depth in the uppermost 8 cm was almost ideally exponential ($R^2 = 0.97$), thus the CF:CS model provided reliable results for the last 130 years. Unfortunately, only one radiocarbon date was available (32 cm, $^{14}$C1200 ± 30BP) due to lack of terrestrial macrofossils preserved in the sediments. Therefore, the simple linear extrapolation based on mean sediment accumulation rate from the CF:CS model was used to establish the primary time scale down core. Comparison to the radiocarbon date showed discrepancy of ca. 150 years (Fig. SI 1) and suggested underestimation of the sediment age. However, verification with pollen-based chronostratigraphic horizon, i.e. sharp decline in human indicators related to the plague of Black Death that peaked in Norway around 1350 CE killing 60% of the human population (Aberth, 2001), showed that the extrapolation based on $^{210}$Pb dating is more reliable than the radiocarbon date, which could be too old. Radiocarbon dates of terrestrial macrofossils indicating older ages can be explained by depositional lags and redeposition of sediment (Bonk et al., 2015).

### 3.2. Chironomidae species composition

Three chironomid zones were identified by the cluster analysis. Chironomid zone 1, between ~1000 and 1370 CE (sediment depth 34–22 cm), was dominated by *H. maeaeri*-type. Procladius, which was frequent in the bottom part of the core decreased towards the end of the zone. In zone 2, between ~1370 and 1850 CE (21–9 cm), *H. maeaeri*-type first decreased but dominated the assemblages afterwards. *H. marcidus*-type and *H. grimshawi*-type were common in the beginning of the zone, whereas *Procladius* increased towards the end of the zone. In zone 3, between ~1850 CE and present (8–1 cm), *H. maeaeri*-type significantly decreased, whereas taxa such as *P. sordidellus*-type, *Stempellinella*, *H. marcidus*-type and *Heterotanytarsus* increased.

### 3.3. Temperature reconstruction

All fossil samples had good modern analogues in the chironomid-based temperature training set based on the MAT. The reconstructed temperatures correlated with PCA axis 1 scores ($r = 0.63$, $r^2 = 0.40$, $p < 0.001$) but not with axis 2 scores ($r = 0.07$, $r^2 = 0.00$, $p < 0.697$). The offset between reconstructed temperature for the topmost sediment sample (2003–2012 CE) and instrumentally measured mean temperature for this period (gridded temperature data based on instrumentally measured temperatures from the Norwegian Meteorological Institute) was 0.2 °C, hence clearly remaining within the sample-specific error estimate.

The reconstructed mean July temperature during the last 1000 years in the Lake Atnsjøen region was on average 11 °C, the maximum values reached 13 °C, whereas the minimum dropped under 9 °C. Four cooler periods were observed: 1050–1150, 1270–1370, 1420–1470 and 1550–1800 CE (Fig. 3).

Between 1000 and 1550 CE warm and cold episodes of 100–150 years duration intermingled (Fig. 3). During cold periods mean July temperature dropped to 9.5 °C, and during warm periods the temperatures increased up to 12–13.5 °C. At the second half of the Millennium, around 1550 CE, a cooler climate prevailed. The mean July temperature dropped to 8.5 °C. From 1800 CE climate warming began and temperatures increased to 12.5 °C (Fig. 3). The recent warming trend visible in the reconstructed temperature curve agrees well with the instrumentally measured temperatures presented as mean July, June to October and annual temperatures (Figs. 2 and 3).

### 3.4. Vegetation changes

According to percentage of pollen grains of trees (AP) and herbs (NAP), two local pollen assemblage zones were separated (Atn-1 and Atn-2). Atn-1 was characterized by the three subzones: a (1000–1250 CE) - with stable share of pine, birch and juniper and visible share of pollen grains and the numerous leafs of *Betula nana*; in subzone b (1250–1450 CE) pollen grains of *Pinus* increased (above 40% with max. 49%) and pollen from *Betula nana* decreased parallel to disappearing of *Betula nana* macro remains in the sediments; in the third subzone - c (1450–1920 CE) growth of percentage of *Juniperus* is visible (with the max. 11%). In the second part of subzone a and c, the decline of *Isoestes lacustris* spores was observed. The share of pollen of pine and birch is stable (about 30–35% each). In the top of LPAZ Atn-2 (from 1920 to 2012 CE) juniper decline and pine again increase (over 45 to 58%) (Fig. 4).

### 3.5. Cladocera and sedimentary TOC

Total Cladocera flux, bosminid sex ratio, and TOC showed several fluctuations during the last 1000 years (Fig. 6). Generally, the total Cladocera and TOC fluxes decreased during cool periods, except for the first cooling in 1050–1150 CE, when only the TOC flux declined. The lowest values of both proxies occurred between 1550 and 1800 CE when the total Cladocera flux dropped to 25 specimens/cm$^2$/yr and TOC flux to

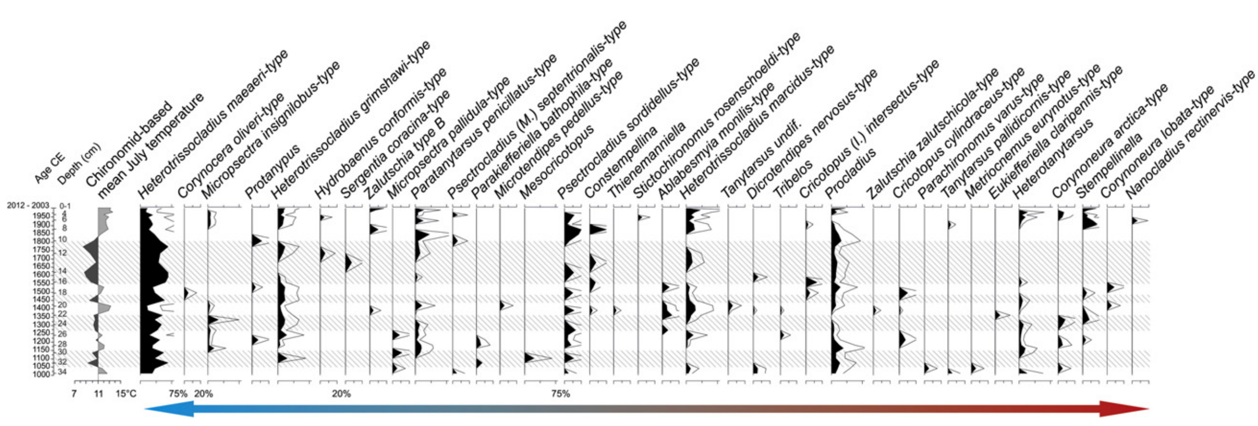


**Fig. 3.** Chironomidae percentage diagram and chironomid-inferred mean July temperature reconstruction from Lake Atnsjøen; Norway. The Chironomidae taxa in the diagram are ordered according to the beta coefficient values calculated in the modern temperature calibration dataset. The species located on the left side of the diagram are cold stenotherms and on the right are the warm indicatorors. Highlighted areas correspond to periods of cooling.

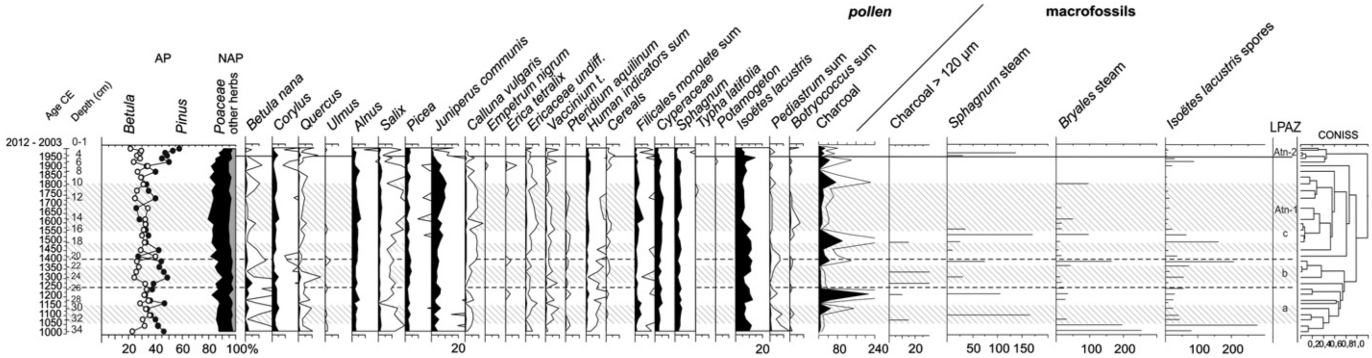


**Fig. 4.** Simplified pollen (%) and macrofossil (counts) diagram of selected taxa from Lake Atnsjøen; Norway. Highlighted areas correspond to periods of climate cooling.

0.06 mg/cm$^2$/yr. Both proxies had higher values in distinguished warmer periods around 1000, 1150–1250, 1400, 1500, and 1800–2012 CE. Since 1900 CE, Cladocera and TOC fluxes began to increase abruptly and reached their maximum values in the second half of the 20th century (Fig. 6): 425 specimens/cm$^2$/yr and 0.9 mg/cm$^2$/yr respectively. The bosminid sex ratio clearly responded to climate cooling between 1700 and 1800 CE, when it reached the highest values 0.6 (Fig. 6).

## 4. Discussion

### 4.1. Climate fluctuation and forcing

The temperature reconstruction was established using the transfer-function approach (Fig. 3). All fossil samples (Fig. 3) had good modern analogues in the chironomid-based temperature training set (Luoto et al., 2009) which confirms the suitability of the training set to the study site. Moreover, since the reconstructed temperatures correlated with PCA axis 1 scores but not with axis 2 scores, it suggests that the variability of the chironomid communities in the core studied was primarily explained by the long-term temperature changes. We found neither indicators of low oxygen conditions nor increased nutrient status throughout the record (Brooks et al., 2007). All the taxa identified were typical of deep, oligotrophic waters. These observations argue against water depth and trophic state control on the chrironomids in Lake Atna during the last Millennium.

Furthermore, the reconstructed temperature value for the surface sediment sample (13.2 °C), representative of modern conditions, fitted well with the instrumentally measured recent temperature at the study area (13.4 °C). Based on these tests we consider the temperature reconstruction reliable.

The beginning of the record (1000 CE) showed elevated temperatures with mean July temperature around 12.6 °C, which is 1.6 °C higher from the millennial mean 11 °C. The other studies from Scandinavia also indicate that the early 11th century was warmer than millennial mean temperature (Gunnarson and Linderholm, 2002; Luoto and Helama, 2010; Osborn and Briffa, 2006) (Fig. 5). The warm climate around the 11th century has been reconstructed in studies from the Northern Hemisphere,but also in central Asia, the Arctic and the Pacific region

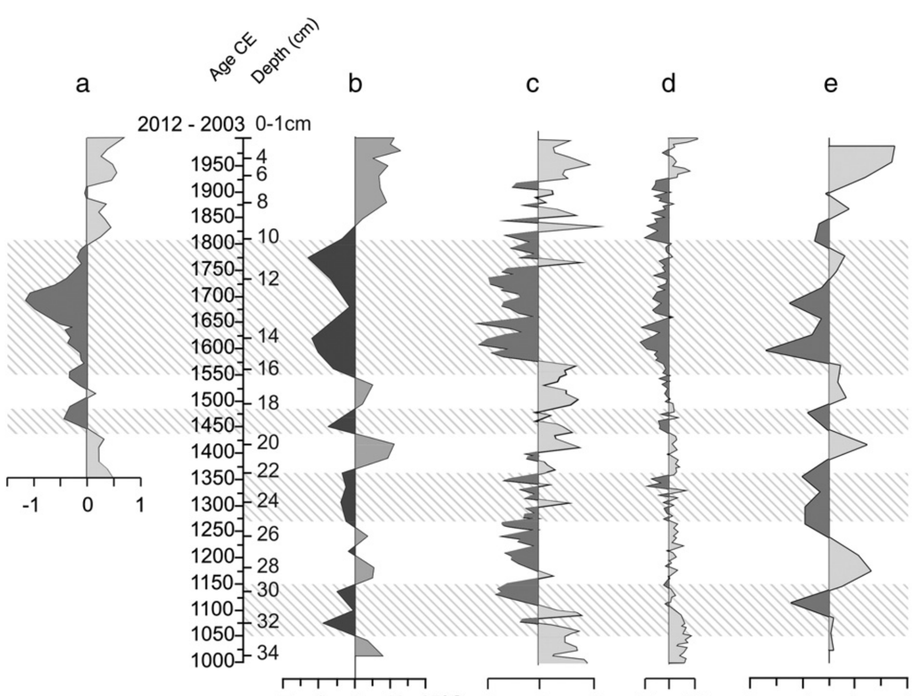


**Fig. 5.** Reconstruction of temperature for the last Millennium (b–e) and last 700 years (a) expressed as an anomaly from the mean (a, c–d) or degree Celsius (b). a) Chironomid-inferred summer air temperature based on combined reconstruction trend from two sites in Finland published by Luoto (2013); b) Chironomid-inferred mean July temperature reconstruction from Lake Atnsjøen; c) tree-ring-based on temperature reconstruction from N Sweden (Torneträsk) published by Osborn and Briffa (2006); d) decadal surface temperature reconstruction averaged over the North Atlantic region based on Mann et al. (2009); e) multi-proxy based temperature reconstruction for Europe published by PAGES 2k consortium (2013). Highlighted areas correspond to periods of climate cooling.

(PAGES 2k Consortium, 2013). Contrastingly at that time, short cooling was observed in Greenland, North America and NW Russia (Mann et al., 2009; Osborn and Briffa, 2006; PAGES 2k Consortium, 2013) suggesting significant regional differences in climate.

At 1050–1150 CE the first of the short-term cooling periods of the last Millennium began and the mean July temperature in the Lake Atnsjøen region dropped to 9.2 °C (Fig. 3). The beginning of this cooling coincided with the Oort solar activity minimum (Fig. 7). The reconstructed climate deterioration agrees very well with temperatures revealed for Europe (PAGES 2k Consortium, 2013) and Finland (Luoto and Helama, 2010), and partly with tree-ring based temperature trends from Northern Sweden (Osborn and Briffa, 2006). There are, however, some differences between the current temperature reconstruction and the tree-ring-based reconstructions from northern Sweden (Fig. 5). These differences can be related to seasonality, as tree growth can reflect springtime conditions rather than July temperatures or the sensitivity of tree-rings to changes in precipitation. However, it is also known that the temperature trends between northern and southern Scandinavia have considerably differed during the past Millennia (Birks and Seppä, 2010), therefore offsets between those records are due to regional climate differences. The climate cooling around 1100 CE has been observed also in Northern America, Russia and Central Asia (Osborn and Briffa, 2006; Wanner et al., 2008), but intrestingly not in Greenland (Osborn and Briffa, 2006). According to presented reconstruction, climate shifted towards warmer conditions during 1150–1250 CE, as mean July temperature raised to 12 °C (Fig. 3). Studies from Finland and Sweden also indicate short climate warming around 1200 CE (Luoto and Helama, 2010; Osborn and Briffa, 2006) (Fig. 5). Other results suggest that in Scandinavia this was a period with temperatures only slightly lower than during the 20th century warming (Gunnarson et al., 2011; Helama et al., 2009). The results presented by PAGES 2k Consortium (2013) indicate climate warming for all continents except Australasia. The above described time interval 1000–1250 CE coincides with the MCA that occurred around 950–1250 CE and was regarded as a generally warmer and drier period (Mann et al., 2009).

In the first half of the last Millennium, presented record revealed two other cold periods: 1270–1370 CE and 1440–1470 CE in which the mean July temperature in Lake Atnsjøen region dropped to 10.2 °C and 9.5 °C respectively (Fig. 3). The beginning of the 1270–1370 CE cooling coincide with Wulf solar activity minimum (Fig. 7) suggesting that the climate was responding to Sun activity. The climate cooling synchronous to this solar minimum had almost global range and it has been recorded from Europe, Arctic, North America and Antarctica (Osborn and Briffa, 2006; PAGES 2k Consortium, 2013) but again not in Greenland (Osborn and Briffa, 2006). The cooling, which has been evidenced also in eastern Finland, has been suggested to represent the initiation of the LIA in Scandinavia (Luoto and Helama, 2010). However, it should be noted that the reconstructed temperature changes at that time are relatively small and comparable to the prediction error of the model (0.72 °C). Moreover in the time between 1000 and 1370 CE, the initial of chironomid zone 1, though dominated by cold-indicating *H. maeaeri*-type, but also characterized by chironomids typical of warm climate conditions (e.g. *Procladius*) (Luoto, 2009).

The beginning of the 1440–1470 CE cold period is synchronous to the pronounce negative NAO phase (Trouet et al., 2009). Other significant phenomena that could have influenced the climate at this time include the Spörer solar activity minimum (Fig. 7), and the large volcanic eruption in 1452 CE that destroyed the Kuwae island (the present territory of Republic of Vanuatu). The climatic impact of this eruption was worldwide, and resulted for example in limited tree growth reported from Finland, Great Britain, France, and China (Witter and Self, 2006). At the regional scale, the 1450 CE cooling has been very clearly reflected in paleoclimate results from southern and eastern Finland (Helama et al., 2009; Luoto, 2013). The climate deterioration at 1440–1470 CE appears to have had a global range and it has been recorded in the results from Europe (Fig. 5), Asia, Arctic, North America, Antarctica and South America (Mann et al., 2009; Osborn and Briffa, 2006; PAGES 2k Consortium, 2013). This cooling is within chironomid zone 2, that lasted between 1370 and 1850 CE. The cold water taxon *H. maeari*-type reached its highest abundances and at the same time, warm-preferring taxa such as *Heterotanytasrus* disappeared from the assemblages. This suggests, that the climate conditions deteriorated already at the end of 15th century, which corresponds to timing of the LIA revealed in other sited in Scandinavia (Luoto et al., 2008; Luoto and Helama, 2010; Rantala et al., 2016).

During 1550–1800 CE, after a short warmer period, a distinct cooling was recorded in the Lake Atnsjøen sediments. Mean July temperature dropped to 9.5–8.5 °C in coldest period, which, according to presented record, is almost 2.5 °C lower than the 1000 yr mean (Fig. 3). The time period of 1600–1700 CE is regarded as the coldest phase of the LIA (Luoto, 2013; Luoto et al., 2008; Luoto and Helama, 2010; Mann et al., 2009). This cooling in Lake Atnsjøen consists of two cold phases divided by short warming around 1650 CE. The beginning of the first phase coincides with changes in circulation in the North Atlantic region as it is synchronous to the beginning of the period with prevailing negative values of NAO index (Trouet et al., 2009). The beginning of the second phase, that started shortly after 1700 CE, is synchronous with the culmination of Maunder solar minimum (Bard et al., 2011). Maunder solar minimum caused a very deep negative NAO index phase (Shindell et al., 2001), which consecutively lead to significant drop in the reconstructed temperature (Fig. 7). The two phased cooling of the LIA is visible also in other reconstruction from Sweden, Finland, North Atlantic region and Europe (Fig. 5) (Luoto et al., 2008; Mann et al., 2009; Osborn and Briffa, 2006; PAGES 2k Consortium, 2013), and in all those reconstruction short warming occurred around 1650 CE. Apparently it occurred slightly earlier in the records from Asia, and was completely absent in records form other continents (Osborn and Briffa, 2006; PAGES 2k Consortium, 2013).

The temperature reconstruction from Lake Atnsjøen indicates that recent and ongoing climate warming began already in 1800 CE following the LIA. Temperatures increased very fast, from 8.5 to 12.8 °C during the first 75 years, but in the 20th century the increase became less pronounced. The time of warming falls in the uppermost chironomid zone which includes a noticeable decrease in the cold adapted *H. maeaeri*-type and a simultaneous increase in warm adapted taxa such as *P. sordidellus*-type, *Stempellinella*, *H. marcidus*-type and *Heterotanytarsus*. Again, the assemblage changes appeared not to be related to a specific limnological change since all the taxa are typical for oligotrophic lakes (Luoto, 2011) suggesting that the assemblage changes reflect climate oscillations.

The warming at the beginning of 19th century in the region of Lake Atnsjøen coincides with a reconstruction from Southern Finland (Luoto, 2013), and a record from Northern Sweden (Osborn and Briffa, 2006). Its onset correlates with the positive NAO index and increased solar activity (Fig. 7). Some authors found that solar change contributed ~60% of the temperature rise since pre-industrial times (Scafetta, 2009; Scafetta and West, 2007). However the anthropogenic impact on the climate is regarded as most important driver of the climate warming in 20th century (Jones and Mann, 2004; Karl and Trenberth, 2003; Lockwood, 2012). It is caused the increased emission of greenhouse gases to the atmosphere. Since 18th century the carbon dioxide concentration increased 31% (Karl and Trenberth, 2003). This is mainly due to industrial activity and the burning of fossil fuels (Elias and Mock, 2013).

In summary, the temperature reconstruction conducted from the sediment record of Lake Atnsjøen suggests, for the first time in this part of Scandinavia, that temperature changes during the first 500 yrs of the last Millennium coincide with solar activity, thought it must be stressed that the 1270–1370 CE cooling was only minor. The cold and warm episodes of similar duration of approximately 100 years interweaved until 1550 CE when the LIA cooling began. As the onset of the LIA coincides with a negative NAO phase, it is possible that the NAO was one of the main drivers of climate changes at that time in western Scandinavia. The presented temperature reconstruction is consistent with work published by the PAGES 2k Consortium (2013) suggesting only minor differences between the local and regional climate trends. This multiproxy temperature reconstruction for Europe indicates that the warming trend in annual temperature began around 1800 CE, as also previously reported by other studies (Luterbacher et al., 2004; Mann et al., 2009). Reconstructions from Eastern Finland suggest that the cold phase corresponding to the LIA lasted until 1900 CE, after which climate began to warm rapidly (Luoto, 2013; Luoto et al., 2008). This distinct lag in climate warming is most likely related to the continentality gradient, which in Scandinavia increases towards the east significantly influencing the local climate conditions (Engels et al., 2014).

### *4.2. Lake-catchment responses to climate*

The distinguished cooler and warmer periods had significant influence on the lake and catchment functioning. The elevated temperatures at the beginning of the record caused an increase of lake productivity, which is shown as higher values of total Cladocera and TOC fluxes (Fig. 6) (Manca et al., 2007; Meyers and Teranes, 2001) indicating increased primary and secondary production in the lake The favourable climatic conditions caused the development of pine forest in the catchment and high production of pollen grains in the lake surroundings reflected by PAR index (Theuerkauf et al., 2013).

The first of the climate cooling periods noted in presented record at 1050–1150 CE caused a slight decrease in lake productivity, which was expressed by lower TOC flux (Fig. 6) suggesting lower production and accumulation of organic matter. The Cladocera based indices, which in Lake Atnsjøen reflect conditions in the pelagic zone due to dominance of euplanktonic Bosmina, did not alter suggesting that food supply (i.e. phytoplankton abundances) remained stable. However, the littoral zone of the lake was affected by climate at this time, probably due to longer ice-cover, and this is expressed in decrease of dominant

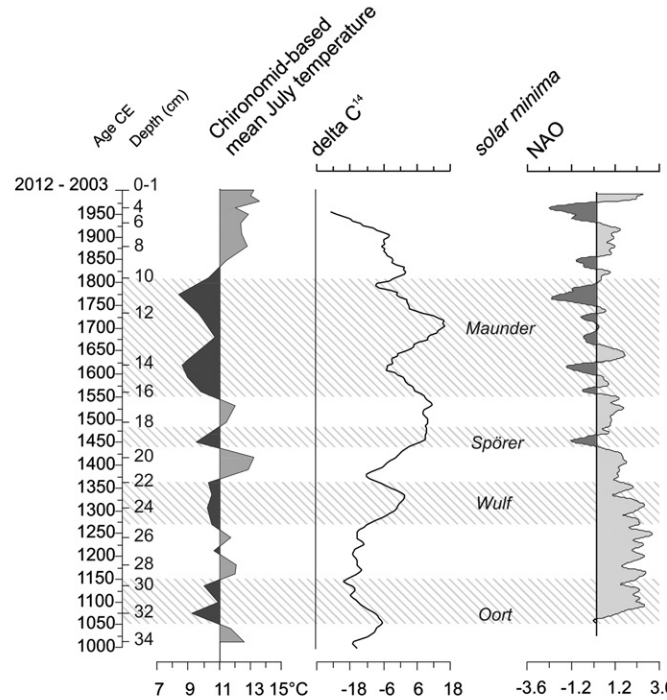


**Fig. 7.** Comparison of chironomid-based temperature reconstruction from Lake Atnsjøen with the results of the delta $^{14}$C curve (Reimer et al., 2013) and anomalies of the NAO index (Trouet et al., 2009). Highlighted areas correspond to periods of climate cooling.

macrophyte *Isoëtes lacustris*. The pollen based PAR of AP indices decreased, and a simultaneous decrease of pine pollen flux suggests that environmental conditions for terrestrial vegetation deteriorated. This is confirmed by the results of a study on pine growth from Central Sweden showing that the mid-12th century was a period characterized by

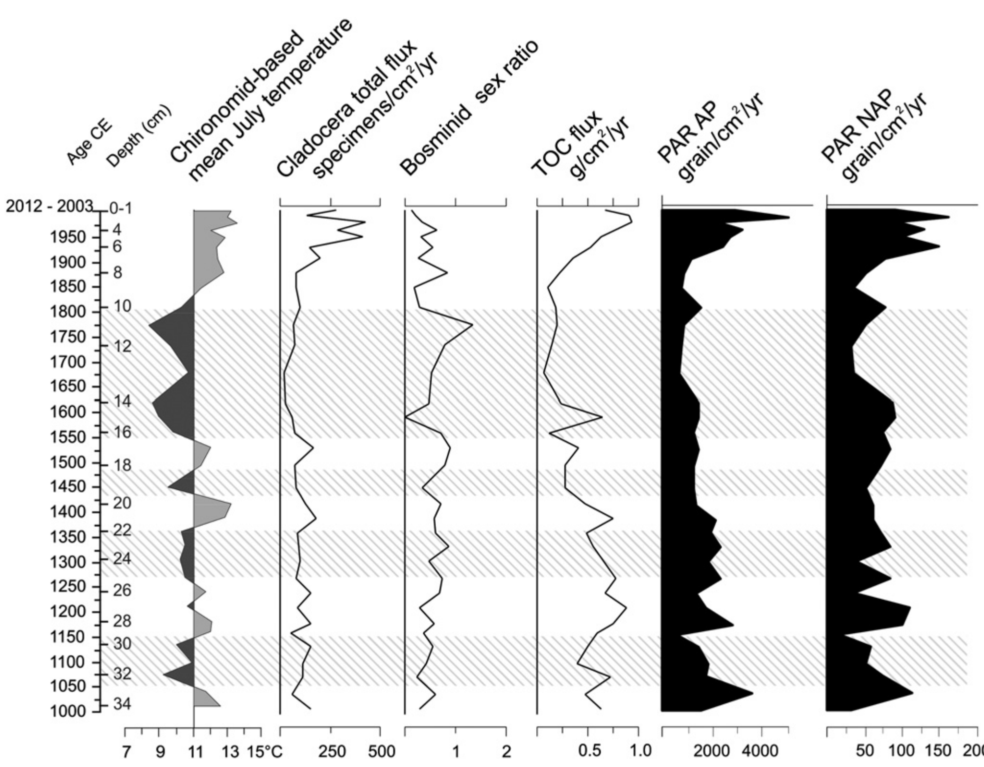


**Fig. 6.** Comparison of chironomid-based temperature reconstruction from Lake Atnsjøen with the selected results obtained from Lake Atnsjøen sediment: Cladocera total flux, bosminid sex ratio, total organic carbon flux (TOC), pollen accumulation rates for arborum (AP) and non-arborum (NAP) pollen flux. Highlighted areas correspond to periods of climate cooling.

low growth and increased pine mortality (Gunnarson and Linderholm, 2002). Another study revealed that the year 1130 CE was the coldest during the last 2000 yrs in this region (Esper et al., 2012).

At 1150–1250 CE, climate warming induced development of forest, which is expressed as higher values of AP and PAR (Theuerkauf et al., 2013). The high number of charcoal in the sediment (Fig. 4) may indicate that summers were dry, as vegetation became vulnerable to fires that occurred more frequently (Drobyshev et al., 2015). On the other hand, charcoal might be an indication of anthropogenic fires connected with slash-and-burn agriculture (Bowman et al., 2011). A slight increase of human activity in the region is suggested by the occurrence of pollen of human indicators and cereals. The lake productivity, reflected by higher TOC and Cladocera flux also slightly increased. Bosminid sex ratio in this time indicate a stable length of the open-water season (Fig. 6) (Nevalainen et al., 2012).

The two other short lasting cooling periods that followed, at 1270–1370 CE and 1440–1470 CE impacted the lake and catchment differently. During the first one, 1270–1370 CE, they were less affected, lake productivity only slightly decreased as suggested by a Cladocera and TOC total flux (Fig. 6). This was probably due to relatively small temperature decrease at that time, only slightly larger than the prediction error of the chironomid-based model. The vegetation in the catchment did not respond, which was corroborated by unchanged values of pollen based indices did not change. The second cold period, 1440–1470 CE was stronger. The temperatures fell by 3 °C which had a more severe influence on lake productivity, indicated by a decrease of TOC flux and to lesser extent by total Cladocera flux. Also the PAR decreased significantly suggesting that this event had a greater impact on the catchment vegetation than during the preceding cooling. The share of pine decreased and remained lower until 1850 CE. Pine pollen in lake sediments reflects not only local but also regional trends, therefore, the decrease of its share provides evidence for deteriorating environmental condition for pines at a broader scale (Scandinavia). The decrease of thermophilous trees such as oak (*Quercus*) started earlier, around 1275 CE, as revealed by studies conducted in Sweden (van der Linden and van Geel, 2006).

During 1550–1800 CE, after a short warmer period, a long lasting cooling marked the Lake Atnsjøen record. This cooling is connected with the coldest phase of the LIA. It triggered the most distinct changes in Lake Atnsjøen and its catchment environment during the last Millennium. From 1550 CE lake productivity strongly decreased as indicated by the lower values of the Cladocera and TOC fluxes. The vegetation in the lake and its catchment was strongly reduced and the dominant littoral plant in the lake, *Isoëtes lacustris*, decreased. However the pollen accumulation rate of AP and NAP reached lowest values throughout the last Millennium around 1650 CE, with a 100 years delay compared to the initiation of the cold phase indicated by aquatic proxies. During 1700–1800 CE conditions for pelagic cladoceran species deteriorated and open-water season was likely reduced, which was indicated by a marked increase in bosminid sex ratio.

The temperature reconstruction from the Lake Atnsjøen record indicates that the ongoing climate warming began already in 1800 CE following the LIA (Fig. 6). However, the indices of lake environment suggest that productivity began to increase with a slight delay at the second half of the 19th century with most prominent increase from 1900 CE onwards. The catchment vegetation showed a similar response, as pine became the dominant tree from 1850 CE onwards, but oak pollen and PAR increased most visibly from the beginning of the 20th century.

## 5. Conclusions

The chironomid-based temperature reconstruction from Lake Atnsjøen in Eastern Norway with mean resolution of 30 years provided evidence that large-scale processes, such as the NAO fluctuations and solar activity modified local climate, and subsequently affected lakes functioning. The three minor cooling periods were reconstructed in the first half of the Millennium: 1050–1150, 1270–1370, 1440–1470 CE, that coincide with solar activity minima: Oort, Wulf, and Spörer respectively. Furthermore, a two peaked cooling period in the second half of the Millennium was identified that coincided with the LIA. These changes co-occurred with the prevailing negative NAO index.

The main late Holocene climate events, namely the MCA, the LIA and were linked to the present climate warming, with other regional records. The temperatures increased at the study site at the 19th century, whereas in the eastern parts of Scandinavia the warming began at the beginning of 20th century. This highlights the need to establish more high resolution climate records to reveal differences in local long-term climate trends, and to identify the forcing mechanisms behind them.

Presented results showed strong impact of the climate on the lake ecosystem and catchment vegetation. Human activity was unimportant factor for the lake ecosystem until the 20th century. All reconstructed coolings resulted in decreases productivity and also a response in surrounding vegetation. The aquatic environment was influenced most pronouncedly during the LIA cooling when biological activity in the lake was strongly limited in the open water zone as well as in the littoral, likely through reduced water temperatures and increased length of the ice-cover period.

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