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Last nine-thousand years of temperature variability in Northern Europe

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Abstract

The threat of future global warming has generated a major interest in quantifying past climate variability on centennial and millennial time-scales. However, palaeoclimatological records are often noisy and arguments about past variability are only possible if they are based on reproducible features in several reliably dated datasets. Here we focus on the last 9000 years, explore the results of 35 Holocene pollen-based July mean and annual mean temperature reconstructions from Northern Europe by stacking them to create summary curves, and compare them with a high-resolution, summary chironomid-based temperature record and other independent palaeoclimate records. The stacked records show that the "Holocene Thermal Maximum" in the region dates to 8000 to 4800 cal yr BP and that the "8.2 event" and the "Little Ice Age" at 500–100 cal yr BP are the clearest cold episodes during the Holocene. In addition, a more detailed analysis of the last 5000 years pinpoints centennial-scale climate variability with cold anomalies at 3800–3000 and 500–100 cal yr BP, a long, warmer period around 2000 cal yr BP, and a marked warming since the mid 19th century. The colder (warmer) anomalies are associated with increased (decreased) humidity over the Northern European mainland, consistent with the modern high correlation between cold (warm) and humid (dry) modes of summer weather in the region. A comparison with the key proxy records reflecting the main forcing factors does not support the hypothesis that solar variability is the cause of the late-Holocene centennial-scale temperature changes. We suggest that the reconstructed anomalies are typical of Northern Europe and their occurrence may be related to the oceanic and atmospheric circulation variability in the North Atlantic–North-European region.

1 Introduction

The climate of Northern Europe is characterized by high multi-scale variability, related to the changing modes and intensities of the atmospheric and oceanic circulation processes (Philipp et al., 2007; Jones and Lister, 2009). It is here that Andersson (1902, 1909), using predominantly fossil plant evidence, laid the foundations of our modern understanding of the general features of post-glacial climatic trends, including the concept of the early- to mid-Holocene warm period, termed here the Holocene Thermal Maximum (HTM), followed by late-Holocene or neoglacial cooling characterized by historically-documented excursions such as the Medieval Warm Period (MWP) or the Little Ice Age (LIA) (Lamb, 1982). As the climate conditions during the Holocene, including the HTM and the neoglacial cooling, provide a reference for the modelled and predicted future climate changes at high latitudes, it is of great importance to understand better the fundamental nature of Holocene temperature variability and its links to external forcing factors, atmospheric and oceanic processes, and feedback mechanisms (Steig, 1999; Kaufman et al., 2004; Rimbu et al., 2004; Seppä et al., 2005; Jansen et al., 2007; Bakke et al., 2008; Beer and van Geel 2008; Wanner et al., 2008).

In the North-European mainland, especially in the lowlands east of the Scandes Mountains, the biological proxies preserved in lake sediments provide the most available and important source for quantitative palaeoclimatological investigations. The most significant recent technical and conceptual advances in using fossil evidence for climate reconstructions in continental regions include the development of robust and realistic quantitative reconstruction techniques (Birks, 1998, 2003), consistently designed, regionally-restricted calibration sets for the development of more reliable organism-based multivariate transfer functions (Birks, 2003; Seppä et al., 2004), and the comparative use of fossil-based reconstructions with independent physical and chemical proxy techniques (Lotter et al., 2002; Seppä et al., 2005). Due to these advances it is possible to produce numerical climate reconstructions than can be used to test palaeoclimatic hypotheses based on climate model simulations, hence offering the possibility of using model-data comparisons for evaluating the relative roles of different climatic forcing factors and feedback responses as drivers of Holocene climatic change (TEMPO, 1996; Prentice et al., 1998; Crucifix et al., 2002; Bonfils et al., 2004; Renssen et al., 2005, 2009). In addition, due to the improved chronological control

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and increased time resolution of the reconstructions it is becoming possible to identify and evaluate statistically the occurrence of centennial-scale warmer and colder periods and to compare reliably the details of the continental palaeoclimatic records with those from marine and ice-core records and the output of climate simulations focusing on fine-scale Holocene variability (Renssen et al., 2006; Jongma et al. 2007).

During the last decade intense efforts have taken place in Northern Europe to create, expand, and improve organism-based calibration models and to produce new quantitative reconstructions so as to increase the accuracy and spatial coverage of the palaeoclimate records. Here we summarize the results of pollen-based Holocene temperature reconstructions along two transects in Northern Europe, ranging from the Norwegian Atlantic coast to 26° E in Estonia and Finland and from 57° N in Southern Fennoscandia to 70° N in the boreal-arctic boundary in Northern Fennoscandia. These transects are designed to allow us to investigate regional patterns in climate history related to the pronounced south-north temperature gradient and west-east oceanicity-continentality gradient of Northern Europe (Giesecke et al., 2008). In addition, inclusion of several temporally-detailed and consistently-generated palaeoclimatic reconstructions provides an opportunity to test the hypotheses about the broad-scale variability of the Holocene climate. An influential but controversial hypothesis suggests that, in addition to the orbitally-forced secular temperature changes, Holocene climate has been repeatedly punctuated by cold events, occurring at roughly 1500-year intervals (Bond et al., 1997, 2001), representing a damped version of the glacial 1470-yr cyclicity, and being possibly connected to reductions in solar output (Bond et al., 2001). Evidence for these repeated cold events, however, is not present in many of the high-resolution marine and terrestrial climate reconstructions from the North Atlantic and its eastern seaboard and the generality of these records has therefore been questioned and intensively discussed (Seppä and Birks, 2002; Risebrobakken et al., 2003; Schulz et al., 2004; Turney et al., 2005; Jansen et al., 2007; Wanner et al., 2008). The new well-dated, high-resolution data sets presented here provide therefore an opportunity to assess the potential occurrence of the hypothesized cold episodes and cycles in the

North-European continental climate history.

2 Area, material and methods

2.1 Pollen-based temperature reconstructions

We carried out pollen-based quantitative climate reconstructions from 35 pollen stratigraphies obtained from lake sediments (Table 1). All lakes were selected and sampled using consistent criteria (Seppä and Birks, 2001; Seppä et al., 2004). Annual mean temperature (T_{ann}) was reconstructed from 12 lakes located in the lowland east of the Scandes Mountians, in Central Fennoscandia and Estonia, between 57°–62° N, in the gradual boundary between the northern temperate zone and southern boreal zone (Fig. 1). July mean temperature (T_{jul}) was reconstructed from 23 lakes located between 68°–70° N in the North-Fennoscandian tree-line region and in the ecotonal regions in Western and Southern Norway (Fig. 1). The altitudinal (alpine) ecotone in Southern Norway, the latitudinal (arctic) ecotone in Northern Fennoscandia and the boreal-temperate ecotone in Southern Fennoscandia and the Baltic countries are predominantly temperature-controlled and represent suitable settings for using pollen data for investigating long-term temperature changes.

The calibration model used for reconstructing the T_{jul} values consists of 321 modern surface-sediment samples of which 283 from Norway, 11 from Svalbard (Norway), and 27 from Northern Sweden (Seppä and Birks, 2001; Birks et al., unpublished). The model used for T_{ann} reconstructions comprises 113 samples from Finland, 24 samples from Estonia and 36 samples from Sweden (Antonsson et al., 2006). Modern T_{jul} and T_{ann} values were estimated using the 1961–1990 Climate Normals data from grids of nearby meteorological stations in Norway, Sweden, Finland, and Estonia. For more detailed information on site selection, fieldwork and modern climate data, see Seppä and Birks (2001) and Seppä et al. (2004).

Modern pollen-climate transfer functions were developed using weighted-averaging partial least squares (WA-PLS) regression (ter Braak and Juggins, 1993). All terrestrial pollen and spore taxa were used in the transfer function. Their percentages were transformed to square-roots in an attempt to optimize the "signal" to "noise" ratio and to stabilize the variances. WA-PLS was selected because it has been shown in many empirical and several theoretical studies to perform as well as or even better than other regression and calibration procedures commonly used to develop organism-environmental transfer functions (see ter Braak et al., 1993; Birks, 1995, 1998).

The performance of the WA-PLS transfer function models are reported (Table 2) as the root mean square error of prediction (RMSEP), the coefficient of determination (R^2) between observed and predicted values, and the maximum bias (ter Braak and Juggins, 1993), all based on leave-one-out cross-validation or jack-knifing (ter Braak and Juggins, 1993; Birks, 1995). Two-component WA-PLS models were selected (Table 2) on the basis of low RMSEP, low maximum bias, and the smallest number of "useful" components (Birks, 1998). More details of the modern pollen-climate data-sets are given in Seppä and Birks (2001), Seppä et al. (2004), and Antonsson et al. (2006).

The reason for reconstructing T_{jul} for the high-latitude sites in Northern Fennoscandia and the high-altitude sites in Western and Southern Norway and T_{ann} for the lowland sites in Central and Southern Fennoscandia is that in the tree-line sites the growing season is confined to three or four summer months (MJJA) and a vegetation-based proxy such as pollen arguably reflects predominantly summer temperature conditions. This is not the case in the Central and Southern Fennoscandian lowlands, where the growing season is considerably longer, starting often in March or April and continuing to October (Walther and Linderholm., 2006). In addition, winter climatic conditions are important for the distribution and regeneration of many plant species, especially those restricted to the most oceanic parts along the west coast of Fennoscandia (Dahl, 1998; Giesecke et al., 2008). Thus the pollen records represent a mixture of taxa with different temperature requirements in relation to the seasons. Annual mean temperature is thus probably a more appropriate climatic parameter to be reconstructed from pollen data in Southern and Central Fennoscandia than July mean temperature (Seppä et al., 2004).

2.2 Age-depth models

The chronology for 33 of the 35 lake-sediment cores is based in AMS radiocarbon dating. The number of dates per core ranges from 4 to 13 (Table 1). All ages were calibrated to calendar years using CALIB4.3 (Stuiver and Reimer, 1993) or CALIB5.0 (Stuiver et al., 2005) software and INTCAL98 (Stuiver et al., 1998) or INTCAL04 (Reimer et al., 2004) calibration data. For sites 1, 3, 8–23 and 25–35 the age-depth models are based on a mixed effect weighted regression procedure within the framework of generalized additive models (Heegaard, 2003; Heegaard et al., 2005). For sites 4, 6, 7, 10, and 24 the age-depth model was obtained by fitting a second-order polynomial curve (third-order with lake 24) to the calibrated dates. Lakes 2 and 5 are particularly important in the present context due to their high sample resolution (Table 1). Lake 2 is an annually laminated lake and has therefore an accurate chronology for the last 9000 years (Ojala and Tiljander, 2003). Lake 5 is partly annually laminated but the varve chronology is floating. The chronology and age-depth model for the lake were derived by correlating the palaeomagnetic secular variation (PSV) curve with the clear anchor points of the PSV curve of lake 2. The obtained chronology is supported by AMS dates (Veski et al., 2004)

The sediment cores were collected in 1990s and early 2000s and the uppermost 0–1 cm of the sediment is believed to represent the present-day.

3 Results and discussion

3.1 General climatic trends

We first examine the implications of the general temperature trends. Figure 2 portrays the results of the individual T_{ann} and T_{jul} reconstructions for the last 9000 years.

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The trends and their differences can be assessed from the LOESS smoothers fitted to the records. Many sites (for example 1, 3, 5, 6, 8, 10, 11, 12) show that T_{ann} was about 2.0–2.5°C higher than at present during the earlier part of the HTM at 8000 to 6000 cal yr BP. Importantly, this is the same temperature deviation as calculated originally by Andersson (1902) in central Sweden and as later reconstructed by the direct borehole temperature measurements of the GRIP ice-core in Greenland (Dahl-Jensen et al., 1998). T_{jul} values at all sites show a lower temperature deviation during the HTM, the maximum values at 8000–6000 cal yr BP being about 1.5°C higher than at present.

Individual time-series climate records are usually noisy and always include chronological error. To be able to distinguish more reliably the main features we calculated the deviations from the mean for all individual records and stacked them into two records of T_{ann} and T_{jul} (Fig. 3a, b) and combined these two records to provide a "stacked summary curve", which shows the general temperature deviations in Northern Europe (Fig. 3c). The stacked T_{ann} record shows a steadily increasing temperature from 9000 cal yr BP onwards, reaching the maximum Holocene level at 8000 cal yr. The subsequent period of highest T_{ann} values, the HTM, lasted over 3000 years, and corresponds therefore with the classical "post-glacial climatic optimum" (Andersson, 1909). As obvious in the individual records (Fig. 2), the magnitude of the HTM warming in the stacked T_{jul} record is lower than in the T_{ann} record. The HTM does not appear as a multi-millennial period, but T_{jul} is highest at 8000–7000 cal yr and declines then steadily towards the present, thus strongly resembling the summer insolation curve for comparable latitudes (Fig. 3d). The stacked summary curve is understandably a combination of these two curves, with a fairly clear HTM at 8000–4800 cal yr BP.

3.2 Early-and mid-Holocene events

The most conspicuous cold event in our records takes place at about 8300 to 8000 cal yr BP, clearly representing the freshwater-forced North-Atlantic 8.2 ka event (Alley et al., 1997; Alley and Àgústsdóttir, 2005; Wiersma and Renssen, 2006). The cooling is present in many T_{ann} records (especially sites 3, 4, 5, 8, 11), mostly located in the ecotone of the temperate and boreal zones, where thermophilous tree taxa occur near their northern range limit. The high-resolution records from this region show a cooling of about 1.0°C, followed by abrupt, high-magnitude warming of about 2.0°C in less than 50 years (Veski et al., 2004). No or weak evidence for the cooling can be observed in the T_{jul} records obtained from Norway, the northern tree-line region, or in the stacked T_{jul} record (Figs. 2 and 3b). Seppä et al. (2007) discuss this spatial pattern and its possible causes. One factor that may explain the clearer evidence in the south is that both evidence and simulations of the 8.2 ka event suggest that the cooling took place mostly during the winter on the eastern North Atlantic seaboard (Alley and Àgústsdóttir, 2005; Wiersma and Renssen, 2006). In the southern part of our study region the vegetative growth pattern, regeneration and pollen productivity are more sensitive to winter and early spring temperatures than in Northern Fennoscandia. However, the relatively weak evidence for the cooling on the Norwegian west coast and Southern Norway (sites 25–35 in Fig. 2), where a strong cooling is suggested by models (Wiersma and Renssen, 2006), is not fully consistent with this explanation and thus requires further attention in the future.

The stacked T_{ann} record shows some variability during the HTM (Fig. 3a), with colder periods at about 7000 and 5300 cal yr BP. These wiggles are not replicated in the T_{jul} record, suggesting that they may not represent regionally significant climatic events. However, Sommer et al. (2009) showed that the regional extirpation of the European pond turtle, a temperate species intolerant of cold summer, happened in Fennoscandia at about 5500 cal yr BP, probably due to a cold spell. Moreover, evidence for a large regional cooling at 5800–5100 cal yr BP has been reported from the North Atlantic and Central Europe (O'Brien et al., 1995; Oppo et al. 2003; Magny and Haas, 2004; Moros et al., 2004; Vollweiler et al., 2006), and the strong signal in the Greenland glaciochemical proxies may be linked to an enchanced Eurasian high (Mayewski et al., 1997), suggesting that the cooling may have been associated with a decreased strength of the westerly circulation in Northern Europe.

3.3 Late-Holocene variability

The stacked records in Fig. 3 show that the last 5000 years have been characterized by a roughly linear cooling trend. To investigate the potential warmer and colder anomalies embedded in this long-term cooling trend, we detrended the stacked summary record for the last 5000 years by fitting a linear curve. The residuals after detrending are shown in Fig. 4a. This curve is compared with a stacked chironomid-based July mean temperature record from Fennoscandia (Fig. 4b), that provides an independent high-resolution summer temperature curve. These two records are compared with two δ^{18}O curves from lacustrine carbonates (Fig. 4c) and an accurately-dated plant macrofossil-based surface wetness record from Southern Finland (Fig. 4d). The δ^{18}O records reflect predominantly temperature changes, but are connected through evapotranspiration to lake-level and humidity changes (Hammarlund et al., 2003; Seppä et al., 2005), whereas the bog surface wetness records in the Baltic Sea region are probably a more direct proxy for changes in effective precipitation and general humidity (Charman et al. 2004; Väliranta et al., 2007).

All five records show generally comparable main features. Three periods of positive deviations and thus high temperature in relation to the trend date to 5000–4000 cal yr BP, 3000–1000 cal yr BP and to the last about 150 years (100 cal yr BP to about AD 2000). The period at 5000–4000 cal yr BP dates to the end of the HTM and is characterized by high temperature and low humidity. These are the typical climatological features of the end of the HTM particularly in the more continental part of Fennoscandia, where the levels of many of the hydrologically-sensitive lakes fell several metres or dried out at after 8000 cal yr BP until a rise after 4000 cal yr BP (Hyvärinen and Alhonen, 1994; Almquist-Jacobson, 1995; Hammarlund et al., 2003; Korhola et al., 2005; Sohar and Kalm, 2008).

The second warm anomaly at 3000–1000 cal yr BP in the pollen-based record is consistent with the positive deviations in the chironomid-based T_{Jul} record, with the rise of δ^{18}O values in the lacustrine carbonate records, and with increasingly dry conditions in the surface wetness reconstruction. This period, which seems to peak at around 2000 cal yr BP, has not been widely investigated or documented earlier in Northern Europe. In central Europe this period appears as a ca. 2000-year long period of relatively high temperature and low humidity. In the Alps, for example, glaciological evidence supported by archaeological finds suggests a marked alpine glacier retreat peaking at 2100–1800 cal yr BP, reflecting thus warm and dry conditions (Jörin et al., 2006; Grosjean et al., 2007). In general, the warm period can be connected with the Roman Warm Period, dating to around 2000 cal yr BP, and with the Medieval Warm Period at about 1000 cal yr BP (Mann, 2007). These two periods are sometimes separated by a shorter colder spell that may have centred on 1500–1400 cal yr BP (Grudd, 2008; Larsen et al., 2008), but this historically documented cold spell ("Dark Age Cold Period") may have been triggered by a volcanic eruption and may be thus too short to be even detected by our stacked data.

It is noteworthy that the MWP cannot be clearly observed in the stacked pollen-based record, nor in the chironomid-based record (Fig. 4). Both records show a generally warm trend with a transition to a colder period starting in the pollen-record already at 1100 cal yr BP and in the chironomid-record at 900 cal yr BP. These features support many earlier investigations according to which the MWP is not reflected as a clear peak in Northern Europe, but rather represents the final centuries of a longer warm period before the onset of cooling at 1000–800 cal yr BP towards the lower temperatures during the LIA (Bradley et al., 2003; Bjune et al., 2009).

The third period with positive temperature deviations dates to last about 150 years. This post-LIA warming has been recorded in many proxy-based reconstructions, for example in the Fennoscandian tree-line region (Weckström et al., 2006; Rosqvist et al., 2007; Bjune et al., 2009). It agrees with historical data about summer and winter temperature trends during the previous five centuries in the region. For example, the longest meteorological records from Sweden show a winter warming since early 1700s (Bergström and Moberg, 2002) and the historical records of ice break-up dates from the Baltic Sea show a winter and spring warming starting already at 1700s and intensifying from the mid 1800s to the present (Tarand and Nordli, 2001). On the basis of their pollen-based reconstructions from 11 sites in the Fennoscandian tree-line region, Bjune et al. (2009) argued that during the 20th century summers were warmest since about 1000 cal yr BP. The same recent warming pattern can be observed in the present, more extensive T_{jul} reconstruction (Fig. 3b) and in the T_{ann} reconstruction reflecting only central and Southern Fennoscandia and the Baltic region (Fig. 3a). The warming that began in the 1800s and reversed the long-term cooling trend has therefore been a large-scale phenomenon in Northern Europe, and probably even in the whole circum-arctic region (Kaufman et al., 2009).

Two colder anomalies can be identified during the last 5000 years, dating to 3800–3000 cal yr BP and to 500–100 cal yr BP. Many records from Northern Europe give evidence of the onset of a cooling trend at about 4500–4000 year cal yr BP, but few previous studies emphatically identify a colder anomaly at 3800–3000 cal yr BP. Some proxy records, for example from Northern Sweden (Rosén et al., 2002) and Finland (Ojala and Tiljander, 2003), suggest a colder period around 3500 cal yr BP and high-resolution sedimentary analyses focusing on the HTM–neoglaciation transition in Southern Sweden pinpoint a cold period at 4000–3500 cal yr, with most severe aquatic response peaking at 3800 cal yr BP (Jessen et al., 2005). This colder event also has equivalence in some records in Northern Europe and North Atlantic (Nesje et al., 2001; Charman et al., 2006). The bog surface wetness record (Fig. 4d) shows that the lower temperature was associated with increased humidity at 3500–3200 cal yr BP (Väliranta et al., 2007). This is consistent with the evidence of Rundgren (2008) who interpreted combined peat-stratigraphical records in Sweden to reflect particularly moist condition, a "wet-shift" peaking at 3300 cal yr.

The last cold anomaly at 500–100 cal yr BP corresponds with the LIA, the most frequently identified cold period in proxy records from Northern Europe. In general, our reconstructed timing and magnitude of the LIA agrees with the results of the more detailed investigations based on dendrochronological data from Northern Fennoscandia (Grudd, 2008) and with the peak of the LIA in Europe, dated from late 1500s to early

1800s (Bradley and Jones, 1993; Moberg et al., 2005). This agreement is noteworthy because it shows that despite the human influence on vegetation composition and land-cover the pollen-based records still capture the main climate trends in the ecotonal areas. This may be partly due to the direct influence of climate on pollen productivity near species distribution limits (Hicks, 1999; Seppä et al., 2007) and partly because most of our sites have been selected from such settings where human influence has been less intense than in the more densely inhabited and cultivated regions. This is particularly true for the sites located in Northern Fennoscandia where the evidence for LIA is clearest (Bjune et al., 2009).

3.4 Forcing factors

A remarkable feature in the climate variability during the last 5000 years is the consistency between the proxies reflecting temperature and humidity. During the last 5000 years the warm anomalies have been associated with dry conditions and cold anomalies with humid conditions. This is undoubtedly partly due to the higher evapotranspiration associated with higher temperature, but is probably partly a result of the nature of the key atmospheric circulation processes in the region. At present in Northern Europe, highest summer temperature anomalies are linked to the anticyclonic circulation type, with the blocking anticyclone as its extreme form, characterized by a long-lived high pressure system centred over Scandinavia, causing weak westerly flow and leading thus to reduced precipitation (Chen and Hellström, 1999; Busuioc et al., 2001; Antonsson et al., 2008; Carrill et al., 2008). Antonsson et al. (2008) suggested that the markedly long warm and dry mid-Holocene period at 8000–5000 cal yr BP in Northern Europe was associated with predominantly anticyclonic summer circulation. The present evidence suggests that this connection between high summer temperature, low humidity and, by inference, anticyclonic circulation may explain even the centennial to multi-centennial- scale climate variability during the late Holocene. The GISP-2 ice-core Cl ion concentration record is often inferred as a proxy for the large-scale circulation type changes over the North-Atlantic-Eurasian region (Fig. 5d). There is

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some correlation, albeit weak, between the Cl record and our temperature record, but this support for the suggested circulation dynamics is tentative at most, especially because the relationship between the ice-core Cl concentration and Eurasian circulation in relatively poorly constrained.

The assessment of the forcing factors behind the inferred climatic and circulation changes is more complicated, but some preliminary assessment can be done by comparing the reconstructed patterns with the proxy records reflecting the magnitude changes in the main forcings. The roughly linear cooling trend during the last 5000 years most likely reflects the high-latitude temperature response to the decline of the summer and annual insolation values (Wanner et al., 2008; Renssen et al., 2009). The suggested deviations from this trend, such as the LIA or the cool period at 3800–3000 cal yr BP, result therefore from the influence of forcing factors other than insolation, with solar irradiation changes and volcanic aerosol, land-cover, and greenhouse gas forcings as the most likely candidates (Wanner et al., 2008). The detrended deviation of the atmospheric $^{14}C/^{12}C$ ratio record is generally interpreted as a proxy for solar irradiance variability during the Holocene (Weber et al., 2004; Beer and van Geel, 2008). In general, the correlation between our temperature records and solar irradiance variability is poor (Fig. 5). Neither the cold anomalies nor the warm anomalies appear to be connected with positive or negative anomalies in the detrended $\delta^{14}C$ data: the LIA maybe an exception to this. The peak of LIA centres in the North-European records to 450–100 cal yr BP, consistent with a positive $\delta^{14}C$ anomaly, and it may have been brought about by the coincidence of low NH orbital forcing during the late-Holocene, with unusually low solar activity and a high number of major volcanic eruptions (Wanner et al., 2008).

4 Conclusions

We combined 35 pollen-based July mean and annual mean reconstructions from Northern Europe to investigate the temperature variability during the last 9000 years.

The records range from the Norwegian Atlantic coast to 26° E in Estonia and Finland and from 57° N in Southern Fennoscandia to 70° N in the North. Most of the records centre on the temperate-boreal boundary in the south, the boreal-arctic boundary in Northern Fennoscandia or the boreal-alpine boundary in the Norwegian mountains. They are therefore sensitively located to capture temperature-driven changes in vegetation composition, vegetative growth patterns, and pollen productivity.

Our results show the well-established pattern of HTM, followed by a roughly linear cooling during the last 5000 years. The coolings at 8200 cal yr BP ("8.2 ka event") and at 500–100 cal yr BP ("LIA") are the most significant abrupt events. The "8.2 ka event" is particularly clear in our stacked T_{ann} record from the Baltic region, whereas the LIA occurs in the whole study region, particularly in the arctic and alpine regions with minimal human interference.

To examine more closely the temperature variability during the last 5000 years we compared our pollen-based detrended temperature record with a stacked chironomid-based July mean temperature record based on the data from seven sites from Fennoscandia. The general features of these two independent records support each other and suggest that, in addition to the cold anomaly during the LIA, another longer late-Holocene cold anomaly dates to 3800–3000 cal yr BP. This anomaly is supported by some high-resolution records, but has not been widely reported earlier. These two cold anomalies are separated by a long warm spell peaking at 2000 cal yr BP, tentatively correlated here with the "Roman Warm Period" reported from Central Europe. The steady late-Holocene cooling trend has been reversed during the last 150 year. This post-LIA warming is consistent with the long-term meteorological and historical records from the region and represents the strongest warming trend since the warming at 8000 cal yr BP, after the 8.2 ka event.

We suggest that the most direct driver of the late-Holocene anomalies has been changes in the dominant atmospheric circulation type. This seems likely in an area, where the modern temperature and precipitation values are highly variable depending on the changing circulation patterns. The anticyclonic circulation type, currently associated with the highest summer temperature, is a strong candidate as the mechanism behind the warm and dry late-Holocene anomalies. A more detailed analysis of the links between the reconstructed temperature patterns, inferred circulation changes, and the key late-Holocene forcing factors, such as the variability in ocean surface temperatures, solar irradiance, aerosols, greenhouse gas concentrations, and more complex combinations of these and other forcings, requires a more coherent analysis involving model experiments and will be a major palaeoclimatological task in the future.

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Table 1. The 35 sites used for the pollenbased temperature reconstructions. The numbers refer to the numbers in Fig. 1.

number	name	alt	long	lat	Modern T_{ann} or T_{jul}	vegetation zone	country	number of samples 9 ka to present	number of ^{14}C dates	reference
1	Laihalampi	137	61°29′ N	26°04′ E	3.9 T_{ann}	Southern Boreal	Finland	122	6	Heikkilä and Seppä (2003)
2	Nautajärvi	104	61°48′ N	24°41′ E	4.0 T_{ann}	Southern Boreal	Finland	240	varved	Ojala et al. (2008)
3	Arapisto	133	60°35′ N	24°05′ E	4.5 T_{ann}	Southern Boreal	Finland	82	7	Sarmaja-Korjonen and Seppä (2007)
4	Kuivajärvi	106	60°48′ N	23°48′ E	4.0 T_{ann}	Southern Boreal	Finland	97	4	Seppä et al. (2009)
5	Rouge	510	57°44′ N	26°45′ E	5.5 T_{ann}	Boreo-Nemoral	Estonia	227	varved, PSV-dated	Seppä et al. (2009)
6	Raigastvere	53	58°35′ N	26°39′ E	5.0 T_{ann}	Boreo-Nemoral	Estonia	90	10	Seppä and Poska (2004)
7	Viitna	74	59°27′ N	26°05′ E	5.0 T_{ann}	North Boreal	Estonia	33	7	Seppä and Poska (2004)
8	Klotjärnen	235	61°49′ N	16°32′ E	4.5 T_{ann}	Southern Boreal	Sweden	104	6	Seppä et al. (2009)
9	Stora Gilltjärnen	172	60°05′ N	15°50′ E	4.6 T_{ann}	Southern Boreal	Sweden	83	8	Antonsson et al. (2006)
10	Lilla Gloppsjön	198	59°50′ N	14°35′ E	5.0 T_{ann}	Southern Boreal	Sweden	73	11	Seppä et al. (2009)
11	Flarken	108	58°33′ N	13°40′ E	5.9 T_{ann}	Boreo-Nemoral	Sweden	91	13	Seppä et al. (2005)
12	Trehörningen	112	58°33′ N	11°36′ E	6.1 T_{ann}	Boreo-Nemoral	Sweden	49	11	Antonsson and Seppä (2007)
13	Litlvatnet	106	68°31′ N	14°52′ E	12 T_{jul}	North Boreal	Norway	35	6	Birks and Peglar, unpublished
14	Myrvatnet	200	68°39′ N	16°23′ E	12.5 T_{jul}	North Boreal	Norway	41	6	Birks and Peglar, unpublished
15	Austerkjosen	135	68°32′ N	17°16′ E	12.8 T_{jul}	North Boreal	Norway	27	6	Birks and Peglar, unpublished
16	Gammelheimenvatnet	290	68°28′ N	17°45′ E	12.8 T_{jul}	Mid/Northern Boreal	Norway	62	8	Birks and Peglar, unpublished
17	Bjørnfjelltjørn	510	68°26′ N	18°04′ E	10.5 T_{jul}	Low Alpine	Norway	46	8	Birks and Peglar, unpublished
18	Vuoskkujavri	390	68°20′ N	19°09′ E	11 T_{jul}	North Boreal	Sweden	49	10	Bigler et al. (2002)
19	Svanåvatnet	243	66°25′ N	14°03′ E	12.1 T_{jul}	Middle Boreal	Norway	61	4	Bjune and Birks (2008)
20	Dalmutladdo	355	69°10′ N	20°43′ E	11.5 T_{jul}	North Boreal	Norway	56	11	Bjune et al. (2004)
21	Toskaljavri	526	69°12′ N	21°28′ E	9.7 T_{jul}	Arcticalpine	Finland	146	8	Seppä and Birks (2002)
22	Tsuolbmajavri	704	68°41′ N	22°05′ E	11 T_{jul}	Arcticalpine	Finland	135	13	Seppä and Birks (2001)
23	Hopseidet	225	70°50′ N	27°43′ E	n.a. T_{jul}	Arcticalpine	Norway	38	4	Seppä et al. (2009); Seppä (1998)
24	KP2	131	68°48′ N	35°19′E	n.a. T_{jul}	North Boreal	Russia	62	8	Seppä et al. (2007)
25	Haugtjern	338	60°50′ N	10°53′ E	14.4 T_{jul}	Southern Boreal	Norway	61	8	Birks and Peglar, unpublished
26	Kinnshaugen	591	61°61′ N	10°22′ E	12.9 T_{jul}	Middle Boreal	Norway	45	8	Birks and Peglar, unpublished
27	Svartvatnet	183	63°21′ N	8°93′ E	12.1 T_{jul}	Southern Boreal	Norway	44	10	Birks and Peglar, unpublished
28	Tiåvatnet	464	63°03′ N	9°25′ E	11.3 T_{jul}	Middle Boreal	Norway	72	8	Birks and Peglar, unpublished
29	Dalane	40	58°15′ N	8°00′ E	14.9 T_{jul}	Nemoral	Norway	47	8	Eide et al. (2006)
30	Flotatjønn	890	59°40′ N	7°33′ E	10.4 T_{jul}	North Boreal	Norway	38	5	Birks and Peglar, unpublished
31	Grostjørna	180	58°32′ N	7°44′ E	15.2 T_{jul}	Southern Boreal	Norway	52	7	Eide et al. (2006)
32	Holebudalen	1144	59°50′ N	7°00′ E	8.2 T_{jul}	Low Alpine	Norway	61	7	Eide et al. (2006)
33	Isbenttjnn	787	59°46′ N	7°26′ E	10.5 T_{jul}	North Boreal	Norway	52	7	Birks and Peglar, unpublished
34	Lille Kjelavatn	1000	59°48′ N	7°15′ E	9.3 T_{jul}	Sub Alpine	Norway	72	6	Eide et al. (2006)
35	Reiarsdalvatnet	245	58°19′ N	7°47′ E	14.3 T_{jul}	Boreo Nemoral	Norway	77	8	Birks and Peglar, unpublished

Table 2. The data and performance statistics, all based on leaveoneout crossvalidation, about the two pollenbased calibration models.

	FES calibration model for T_{ann}	NFS calibration model for T_{jul}
number of sites	173	321
temperature gradient	−4.1 to 7.1°C	3.5–16.4°C
number of taxa	104	183
RMSEP	0.95°C	1.135°C
R^2	0.88	0.77
maximum bias	2.1°C	2.53°C

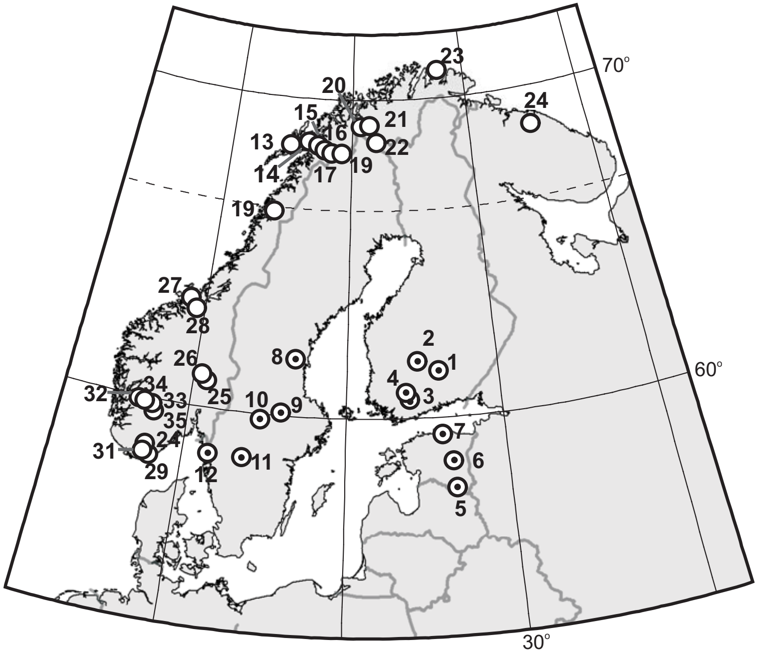

Fig. 1. The location of the sites from where the quantitative pollen-based temperature reconstructions where derived. The filled circles indicate sites with T_{ann} reconstructions and open circles with T_{jul} reconstructions. For the name of the sites, see Table 1.

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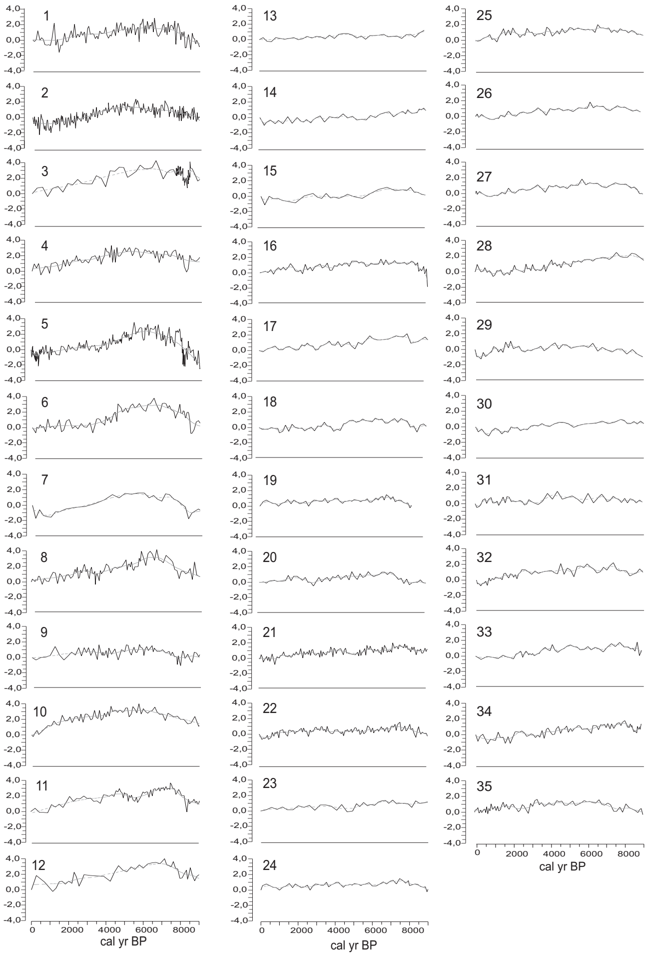

Fig. 2. The individual pollen-based T_{ann} and T_{jul} reconstructions for the last 9000 years. Sites numbered as in Table 1.

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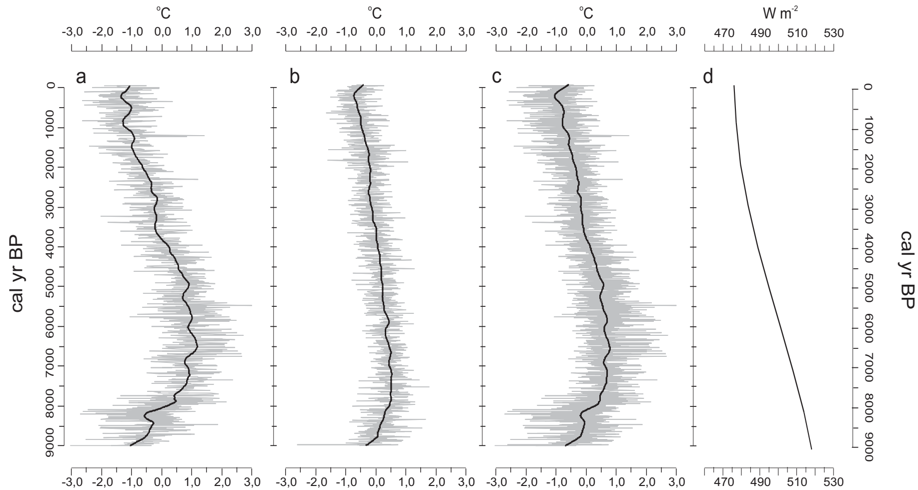

Fig. 3. The North-European pollen-based stacked T_{ann} **(a)** and T_{jul} **(b)** records. The T_{ann} record comprises a total of 1291 reconstructed T_{ann} values (average sample interval of 7.0 years) and the stacked T_{jul} record is based on 1561 values (average sample interval 5.8 years) for the 9000 years. The temperature values are expressed as deviations from the mean. These two records are combined in **(c)** to show "a stacked summary curve" reflecting the North-European temperature variability record, consisting of 2852 values with an average sample interval of 3.2 years. All records are shown with a LOESS smoother with a span of 0.05. **(d)** June insolation at 60° N northern latitude (Berger and Loutre, 1991).

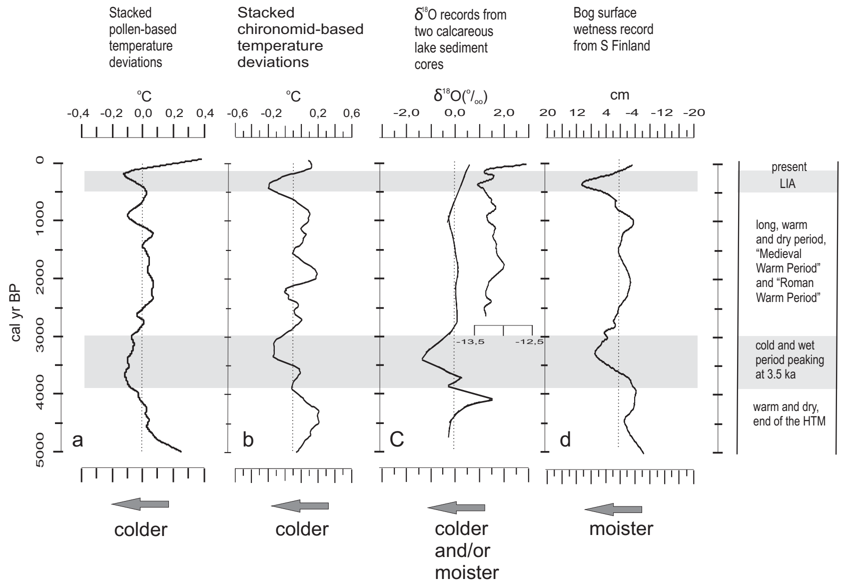

Fig. 4. The temperature variability in Northern Europe during the last 5000 years. **(a)** The pollen-based temperature variability record based on the stacked summary record. The reconstructed trend during the last 5000 was detrended by adding a linear curve and the residuals are shown here, smoothed with a LOESS smoother with span 0.1. **(b)** A chironomid-based July mean temperature variability, as reflected by residuals after detrending by adding a linear curve. The chironomid-based curve is a stacked record, based on six sites in Norway (Velle et al., 2005) and one site, Toskaljavri, in Northern Finland (Seppä et al., 2002), showing the deviations from the mean with a LOESS smoother with span 0.1. **(c)** Two δ^{18}O-based records from lacustrine calcareous sediments, from Lake Igelsjön in Southern Sweden (Hammarlund et al., 2003) (residuals after detrending), and Lake Tibetanus in Northern Sweden (Rosqvist et al., 2007), reaching back to 2600 cal yr BP, **(d)** a general humidity record based on bog surface wetness changes reconstructed quantitatively from plant macrofossil composition in Southern Finland (Väliranta et al., 2007).

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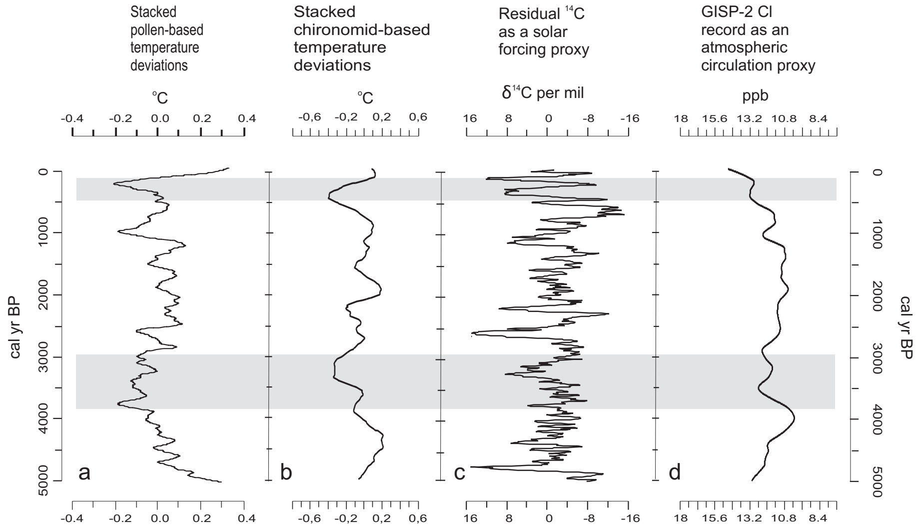

Fig. 5. Comparison of the reconstructed late-Holocene temperature variability with proxies that reflect possible forcings of the high-frequency variability. **(a)** The stacked summary pollen-based temperature variability record, as in Fig. 4a but with LOESS smoother with 0.03. **(b)** The chironomid-based temperature variability record as in Fig. 4b, **(c)** The δ^{14}C residuals as a proxy for solar radiation variability, unsmoothed, **(d)** The GISP-2 Cl ion concentration record as a proxy for the North Atlantic atmospheric circulation pattern variability (Mayewski et al., 1997). Smoothed with a LOESS smoother with span 0.1.