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Holocene annual mean temperature changes in Estonia and their relationship to solar insolation and atmospheric circulation patterns

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Abstract

We reconstructed annual mean temperature (T_{ann}) trends from three radiocarbon-dated Holocene pollen stratigraphies from lake sediments in Estonia, northern Europe. The reconstructions were carried out using a North-European pollen-climate calibration model based on weighted averaging partial least-squares regression. The cross-validated prediction error of the model is 0.89°C and the coefficient of determination between observed modern T_{ann} values and those predicted by the model is 0.88. In the reconstruction, the Holocene thermal maximum (HTM) is distinguishable at 8000–4500 cal yr B.P., with the expansion of thermophilous tree species and T_{ann} on average 2.5°C higher than at present. The pollen-stratigraphical data reflect progressively warmer and drier summers during the HTM. Analogously with the modern decadal-scale climatic variability in North Europe, we interpret this as an indication of increasing climatic continentality due to the intensification of anticyclonic circulation and meridional air flow. Post-HTM cooling started abruptly at around 4500 cal yr B.P. All three reconstructions show a transient (ca. 300 years) cooling of 1.5–2.0°C at 8600–8000 cal yr B.P. We tentatively correlate this cold event with the North-Atlantic "8.2 ka event" at 8400–8000 cal yr B.P. Provided that the 8.2 ka event was caused by freshening of the North-Atlantic surface water, our data provide evidence of the climatic and vegetational responsiveness of the boundary of the temperate and boreal zones to the weakening of the North-Atlantic thermohaline circulation and the zonal energy transport over Europe. No other cold events of comparable magnitude are indicated during the last 8000 years.

Keywords: estonia; annual mean temperature; calibration set; Holocene thermal maximum; 8.2 ka event; pollen stratigraphy

Introduction

The development of climate model simulations have advanced palaeoclimatological research from descriptive reconstructions towards understanding the past atmospheric dynamics and interactions of the components of the global climate system. Past climatic patterns have been simulated with different model configurations with atmosphere-only models (de Noblet et al., 1996; Hall and Valdes, 1997), with coupled atmosphere–ocean general circulation models (CGCMs) (Grassl, 2000) and with earth system models of intermediate complexity (EMICs) combining atmosphere, ocean, and vegetation (Claussen et al., 2002; Crucifix et al., 2002). The CGCMs and EMICs demonstrate strong synergism between the components of the climate system, indicating, for example, that during the mid-Holocene at high latitudes, higher-than-present summer insolation and related high summer temperatures led to a decrease of the planetary albedo due to the expansion of the boreal forest and to an amplification of the warming by the sea-ice albedo feedback (Ganopolski et al., 1998). Model runs are, however, specific to certain points in time, have a coarse geographical resolution, and often over- or underestimate the influence of regional climatological characteristics (Crucifix et al., 2002). Palaeoclimatological reconstructions based on proxy evidence have an important role in generating regionally and temporally more precise palaeoclimatological time-series data which can be used for data–model comparisons in order to validate model simulations and to achieve a more coherent picture of past atmospheric dynamics.

Simultaneously with the development of climate models, the performance of the quantitative palaeoclimatological reconstructions based on biological proxies has been

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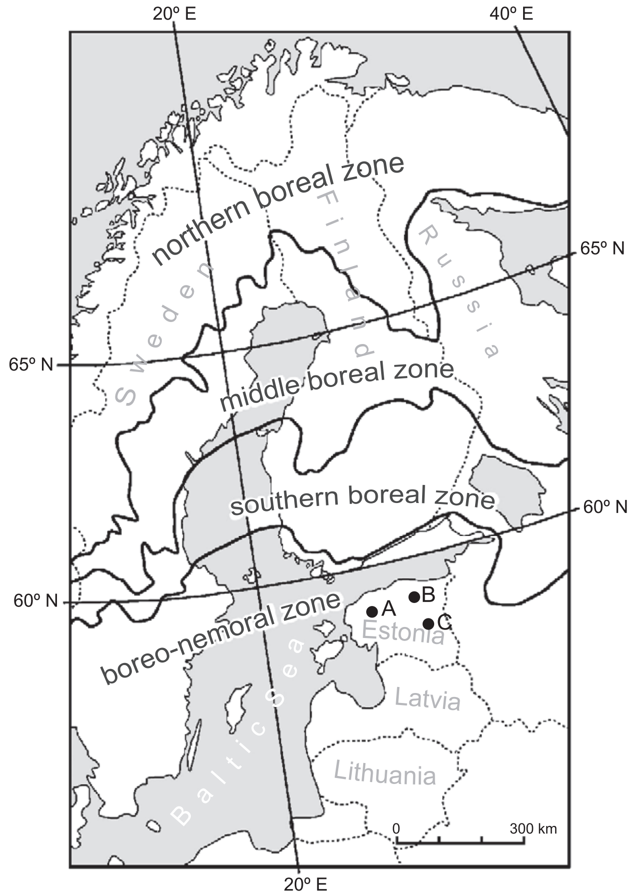

Fig. 1. The location of the study sites in Estonia. A = Lake Ruila, B = Lake Viitna, C = Lake Raigastvere. The vegetation zones of North Europe are according to Ahti et al. (1968).

boosted by progress in the development of numerical inference techniques and computer software (see Birks, 1998). Equally important to this technical development, has been an increase in understanding the significance of critical ecological and biogeographical considerations in the design of Holocene research projects. Of particular importance is that reconstructions must be focused on regions where the palaeorecords are climatologically sensitive and where it is possible to create high-quality modern calibration sets (Birks, 1995). Within the scope of pollen-based climate reconstructions, such regions are where there exists a simple zonal climatic gradient, determined, or strongly influenced, by one or few dominant climatic variables, and, related to this, where there exists an equally clear vegetation zonality (Seppä et al., 2003). These requirements are met in northern Europe where there is a distinct North–South gradient of such key climatic variables as annual mean temperature and the length of the growing season from the boundary between the boreal and arctic climatic zones at ca. 70°N latitude to the northern limit of the temperate zone at ca. 55°N latitude (Johannessen, 1970) with a corresponding change in vegetation (Moen, 1999). The high performance statistics of a recently compiled Finnish–Estonian pollen–climate reconstruction model, based on modern pollen samples along a transect across the boreal zone and corresponding modern climatic data, indicate the potential sensitivity of pollen records in northern Europe as archives of past climate changes (Seppä et al., 2003).

Here we report Holocene annual mean temperature reconstructions based on three pollen stratigraphies in Estonia, northern Europe, and on the Finnish–Estonian pollen–climate reconstruction model of Seppä et al. (2003). Our aim is to produce quantitative annual mean temperature records, to assess the role of past astronomical solar insolation and atmospheric circulation patterns as causal factors for the reconstructed changes, and to contribute to the current discussion of Holocene climate variability in northern Europe over long and short timescales. Estonia is located in the boundary region of the temperate and boreal climatic domains and its climate is transitional between the continental climate of eastern Europe and the oceanic climate of western Europe. The characteristic weather type changes sensitively and rapidly according to changes of dominance between the western zonal flow and continental high-pressure conditions (Johannessen, 1970). Vegetationally, Estonia can be characterized as southern coniferous (Walter and Breckle, 1986), with a dominance of boreal conifers but with locally significant occurrences of temperate deciduous tree species. Due to these climatic and vegetational characteristics, the past vegetation patterns of Estonia and fossil pollen assemblages are likely to have been sensitive to the variability of the major components of the atmospheric circulation over northern Europe and provide a suitable basis for the current work.

Study area

Estonia is located between 59°49′N and 57°30′N and between 21°46′E and 28°13′E (Fig. 1). The geological relief is even, with the highest point 318 m a.s.l. The sedimentary bedrock consists predominantly of Ordovician

Table 1
Information about the three study sites

	Lake Raigastvere	Lake Viitna	Lake Ruila
Latitude N	58°35′	59°27′	59°10′
Longitude E	26°39′	26°05′	24°26′
Altitude	52.5 m a.s.l	74.4 m a.s.l	43.0 m a.s.l.
Basin size	122 ha	16.3 ha	16.6 ha
T_{ann}	5.0°C	5.0°C	5.5°C
T_{july}	16.5°C	16.5°C	16.5°C
T_{jan}	−6.5°C	−5.5°C	−4.5°C
Precipitation	650 mm/yr	650 mm/yr	700 mm/yr
Vegetation	Coniferous forest	Coniferous forest	Coniferous forest

The modern climatological data are based on the Climate Normals period 1961–1990.

Table 2
The statistics of the Finnish–Estonian pollen–climate calibration set

Number of lakes	137
Temperature gradient	−4.7–5.5°C
Temperature range	10.2°C
Mean temperature	1.9°C
Median temperature	2.4°C
Number of taxa	102
r^2 between predicted and modern temperature	0.88
Root mean square error of prediction	0.89°C
Root mean square error of prediction as % of the gradient length	8.8%
Maximum bias	2.13°C

The performance statistics of the set are based on a two-component weighted averaging partial least-squares regression model.

and Silurian carbonaceous rocks in the north and of Devonian sandstone in the south (Viiding, 1995). During the Quaternary Period, Estonia was repeatedly glaciated. The Quaternary deposits, usually less than 5 m thick, cover the bedrock.

Estonia is located at the zonal boundary of the temperate and boreal climates and of the Atlantic and continental sectors. Annual mean temperature varies from ca. 6.0 °C in the west to ca. 4.5°C in the east. Summers are warm, with July mean temperatures around 17.0°C in most parts of the country, and winters are moderately cold, with January mean temperatures ca. −6.5°C in the east and ca. −2.0 to −5.0°C in the west. The winter climate is particularly dependent on the dominant atmospheric circulation mode. During the dominance of the zonal flow, winter temperatures are about 5.0 to −5.0°C, while the intensification of the Eurasian (Siberian) high-pressure often leads to southward extensions of polar air and winter temperatures fall to −20 to −30°C (Johannessen, 1970).

Estonia is located in the hemiboreal vegetation zone. Forests are dominated by pine (*Pinus sylvestris*), birch (*Betula pendula* and *B. pubescens*), and spruce (*Picea abies*), with lesser proportions of alder (*Alnus incana*, *A. glutinosa*) and aspen (*Populus tremula*). Temperate deciduous trees such as oak (*Quercus robur*), hazel (*Corylus avellana*), lime (*Tilia cordata*), elm (*Ulmus glabra*), and ash (*Fraxinus excelsior*) occur infrequently. Due to the ecotonal nature of Estonia, many of the temperate tree species are close to their northern distribution limit.

The pollen stratigraphies on which the climate reconstructions are based were collected from three lakes, Lake

Table 3
Radiocarbon dates for the three study sites

Lab. no	Depth	Dated material	Age ^{14}C yr B.P.	Calibrated age (cal yr B.P.)	2σ interval (cal yr B.P.)
Lake Raigastvere					
Ta-1759	410–420	gyttja	520 ± 70	530	650–480
Ta-1758	510–520	gyttja	1810 ± 80	1740	1900–1540
Ta-1757	610–620	gyttja	3080 ± 90	3310	3480–3010
Ta-1756	710–720	gyttja	3790 ± 60	4170	4380–3970
Ta-1755	810–820	gyttja	4400 ± 80	4950	5300–4840
Ta-1754	910–920	gyttja	5350 ± 80	6160	6300–5940
Ta-1753	1010–1020	gyttja	6310 ± 60	7240	7370–7050
Ta-1752	1110–1120	gyttja	7230 ± 80	8020	8190–7900
Ta-1751	1210–1220	gyttja	7750 ± 100	8510	8770–8370
Ta-1749	1260–1270	gyttja	8180 ± 70	9100	9370–8990
Lake Viitna					
Tln-2141	490–500	gyttja	1650 ± 50	1540	1670–1420
Tln-2140	540–550	gyttja	3760 ± 70	4120	4370–3920
Tln-2143	610–620	gyttja + moss	5360 ± 105	6170	6350–5910
Tln-2144	690–700	coarse detritus gyttja	7320 ± 60	8140	8220–8000
Tln-2145	772–782	gyttja	8610 ± 70	9560	9740–9480
Tln-2146	792–802	gyttja	9200 ± 70	10,340	10,550–10,210
Tln-2147	812–822	gyttja	10,520 ± 85	12,650	12,840–12,340
Lake Ruila					
Tln-2523	145–150	gyttja	2740 ± 125	2810	2970–2750
Tln-2524	165–170	gyttja	3210 ± 85	3430	3630–3250
Tln-2526	185–190	gyttja	3340 ± 85	3570	3800–3370
Tln-2533	215–220	gyttja	3890 ± 85	4320	4430–4180
Tln-2530	270–275	gyttja	4850 ± 70	5600	5640–5560
Tln-2531	292–297	gyttja	6170 ± 90	7010	7280–6770
Tln-2534	315–320	gyttja	6850 ± 70	7670	7720–7580
Tln-2535	348–353	gyttja	8160 ± 45	9080	9260–9010
Tln-2538	380–385	gyttja	9620 ± 115	11,080	11,220–10,590
Tln-2546	438–443	gyttja	10,390 ± 150	12,330	12,900–11,620
Tln-2545	443–453	gyttja	10,390 ± 160	12,330	12,900–11,580

Raigastvere, Lake Ruila, and Lake Viitna (Fig. 1, Table 1). Local descriptions for Lake Ruila are provided by Poska and Saarse (2002) and for Lake Viitna by Saarse et al. (1998). The modern climatic conditions of the lakes, interpolated from the closest meteorological stations, are given in Table 1. The lakes are mostly surrounded by forests, but, as is common in Estonia, there is significant human impact on the vegetation near the lakes, and fields are common.

Methods

For the climate reconstructions, we used a transfer function approach based on a combined Finnish–Estonian pollen–climate calibration set (Table 2). For this set 137 lake-sediment surface samples (ca. top 10 mm) were collected from small to medium-sized lakes in Finland (113 samples) and Estonia (24 samples), and a minimum of 500 terrestrial pollen and spores were analyzed from the surface samples. The samples were collected, prepared, and analyzed in a standardized way (see Seppä et al., 2003).

Modern pollen-zclimate transfer functions for the climate variable of interest were constructed using weighted averaging partial least squares (WA-PLS) regression (ter Braak, 1995; ter Braak and Juggins, 1993). Transfer functions were developed using all terrestrial pollen and spore types, with their percentages transformed to square roots in order to optimize the 'signal-to-noise' ratio and to stabilize the variances. The performance of the resulting reconstruction model was tested with the modern material, using leave-one-out cross-validation (ter Braak and Juggins, 1993), where the modern climate was reconstructed or "predicted" n times using a training set of size $n-1$, omitting the sample from the site for which the modern climate was predicted (see Birks, 1995). In our model, the coefficient of determination (r^2) between observed and predicted annual mean temperatures is 0.88, the root-mean-square error of prediction (RMSEP) is 0.89°C, and the maximum bias is 2.13°C, all based on leave-one-out cross-validation (Table 2). For more details of the calibration set, including a discussion of the potential effect of human impact on vegetation and the pollen-climate calibration model, see Seppä et al. (2003).

The modern pollen–climate transfer functions were applied to the fossil pollen assemblages of the three sediment cores (Fig. 1; Table 1). These pollen analyses were carried out by A. Poska and R. Pirrus No taxonomic harmonization was carried out between the two analysts. Both with the training set and with the sediment cores, pollens were identified to the lowest possible level and all the terrestrial pollen and spore taxa were included in the reconstructions. Percentages of terrestrial pollen and spore taxa were calculated on the basis of their total sum. Pollen diagrams were prepared with TILIA and TILIA.GRAPH (Grimm, 1990). The sample-specific RMSEP for each reconstructed sample was generated by Monte Carlo simulation with the program WAPLS (Birks, 1995; ter Braak and Juggins, 1993).

Chronological control for the three sediment cores was provided by conventional bulk radiocarbon datings (Table 3). Radiocarbon dates were calibrated with CALIB 4.2 program (Stuiver and Reimer, 1993) and INTCAL98 calibration data (Stuiver et al., 1998), using bidecadal tree-ring data set A and method A, and applying ten-sample curve smoothing. Age–depth modeling for each sequence was performed by linear interpolation between the calibrated radiocarbon dates (Fig. 2). A modern age was assumed for the uppermost sediment of the cores. Due to the lack of radiocarbon dates from the basal part of the cores or large confidence limits for the oldest calibrated dates, the chronologies become more unreliable towards the lower parts of the cores (Fig. 2). For this reason, the T_{ann} reconstructions were confined to the last 9000 yr.

Results and discussion

Selection of the climate variable of interest

The selection of the climatic variable(s) is a critical question in pollen-based climate reconstructions. Pollen stratigraphy reflects the composition of past vegetation and the most relevant climatic variables are those that exert

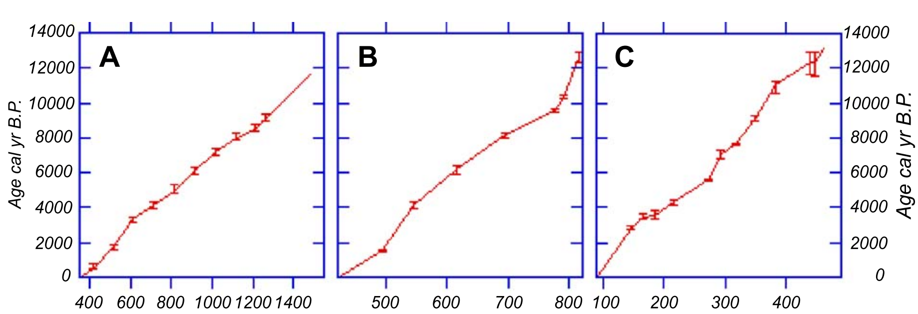

Fig. 2. Age–depth models for the Lake Raigastvere (A), Lake Viitna (B), and Lake Ruila (C) sediment cores. The models are based on linear interpolation between successive calibrated radiocarbon dates (see Table 3). The vertical bars show the 2σ ranges of the calibrated ages.

the strongest influence on the vegetation. These are often different from the climatic variables recorded by instrumental or historical palaeoclimatological time series, such as the 500-year-long historical ice-cover record from Tallinn, Estonia, reflecting mostly winter and spring temperatures (Tarand and Nordli, 2000), or from those reflected by physical and chemical palaeoclimatological proxies such as ice-core borehole temperatures or $\delta^{18}O$ records (Johnsen et al., 2001). In plant ecology, the length of the growing season, the growing degree-days or growing degree-days above 5°C, which takes into account both the length of the growing season and summer warmth, are often suggested as the most important bioclimatic variables (Woodward, 1987; Crucifix et al., 2002). The selection of climate variables is, however, made more complicated by the highly individualistic climatic relationships of different plant species (Dahl, 1998).

It is realistic to accept that no single climatic variable can account for the complete influence of climate on vegetation and that no single or few reconstructed climate variables can reflect the full spectrum of past climate patterns and changes. Thus, in this paper, we have reconstructed annual mean temperature (T_{ann}), an important, simple climatic variable, which can be used for correlations with other climate reconstructions and palaeoclimatic proxy records. T_{ann} has a strong zonal pattern in the study area, aligned with the vegetation zonation, and has high r^2 and low RMSEP in the leave-one-out cross-validation of the pollen–climate calibration set (Table 2) and therefore accounts for a large proportion of the total spectrum of temperature-related climate patterns.

Climate reconstructions: low-frequency trends

The simplified pollen diagrams are shown in Fig. 3 and the three T_{ann} reconstructions in Fig. 4. Due to the high variability between single samples and the low number of samples per cores, it is useful to summarize the Holocene temperature trend by means of regionally averaged departures from modern values (Fig. 5). All three reconstructions suggest similar T_{ann} patterns during the early to mid-Holocene, with T_{ann} of ca. 5.5–6.5°C at 9000–8000 cal yr B.P., rising to the highest Holocene level (Holocene thermal maximum, HTM) at 8000–7500 cal yr B.P. All three data sets indicate remarkably high T_{ann} values during the HTM, with reconstructed T_{ann} 8.0–9.0°C. The regionally averaged data show that at this time T_{ann} was about 2.5°C higher than the reconstructed modern value. This agrees with the results of the GRIP ice-core borehole temperature measurement (Dahl-Jensen et al., 1998) and roughly with the pollen-based T_{ann} reconstruction of Heikkilä and Seppä (2003) from southern Finland, but is ca. 0.5–0.9°C higher than in the reconstructions from the Fennoscandian treeline area, ca. 10° north of Estonia (Seppä and Birks, 2001, 2002).

Importantly, the temperature difference between the HTM and the present in the Estonian reconstructions is also in agreement with the atmosphere–ocean–vegetation (AOV) climate model simulations of Ganopolski et al. (1998). AOV simulations suggest that the difference between mid-Holocene summer (JJA) temperature in relation to the modern summer temperature was ca. 2.5°C in eastern North Europe during the mid-Holocene with a higher anomaly towards the east and lower toward northwestern Europe. Although the climatic variables reconstructed by our model and simulated by the AOV climate model are not strictly the same, these two independent approaches to estimating the amplitude of mid-Holocene warming in relation to modern conditions support each other. This implies that the AOV climate model simulates correctly the feedback responses of the oceanic and vegetational components of the climate system to the external forcing patterns during the mid-Holocene.

A closer inspection of the pollen stratigraphical data suggests significant climate changes during the HTM that are not reflected as T_{ann} variations. During the earlier part of the HTM (9000–8000 cal yr B.P.) the dominant temperate deciduous tree taxa were *Corylus*, *Ulmus*, and *Alnus*, whereas the percentages of *Quercus*, *Tilia*, and *Fraxinus* pollen are low, start to rise at 7000–6000 cal yr B.P., and have their maximum values at 7000–4000 cal yr B.P. (Fig. 3). In northern Europe *Corylus avellana* and *Ulmus* (*glabra*) start to grow early in the spring and demand a long growing season. They are, therefore, slightly oceanic species, whereas *Quercus*, *Tilia*, and *Fraxinus* are the tree species of the North European boreal zone which, in addition to a long-growing season, require the highest mid-summer temperatures for successful generative reproduction (Dahl, 1998; Hintikka, 1963; Pigott, 1981; Pigott and Huntley, 1981; Prentice and Helmisaari, 1991; Skre, 1979). The rise of the pollen values of these taxa implies therefore that there was a change in seasonality at ca. 7000–6000 cal yr B.P. from a more oceanic climate toward a climate characterized by particularly high midsummer temperatures and dry summers, whereas the rise of *Picea* toward the end of the HTM suggests that winters may have become colder and snowier. Consequences of such a climate change should be reflected in the long-term hydrological regime of the region; Saarse et al. (1995) provide evidence for a major lake-level lowering in Estonia at ca. 5000–3500 cal yr B.P., apparently as a result of high summer temperatures and low precipitation.

The late-Holocene climate is characterized by the gradual decrease of T_{ann} from the HTM toward modern conditions and, as indicated by lake-level data (Saarse et al., 1995), by increasing humidity. This is consistent with peat stratigraphy of Estonian raised bogs (Karofeld, 1998). The development of microtopography with hummocks and hollows started ca. 3000 cal yr B.P., probably as a result of lower temperatures, decreased evapotranspiration, and subsequent increase of surface wetness (Karofeld, 1998).

These inferred Holocene climate changes in Estonia are clearly not due only to the astronomical solar insolation changes, as both annual and summer solar insolation values

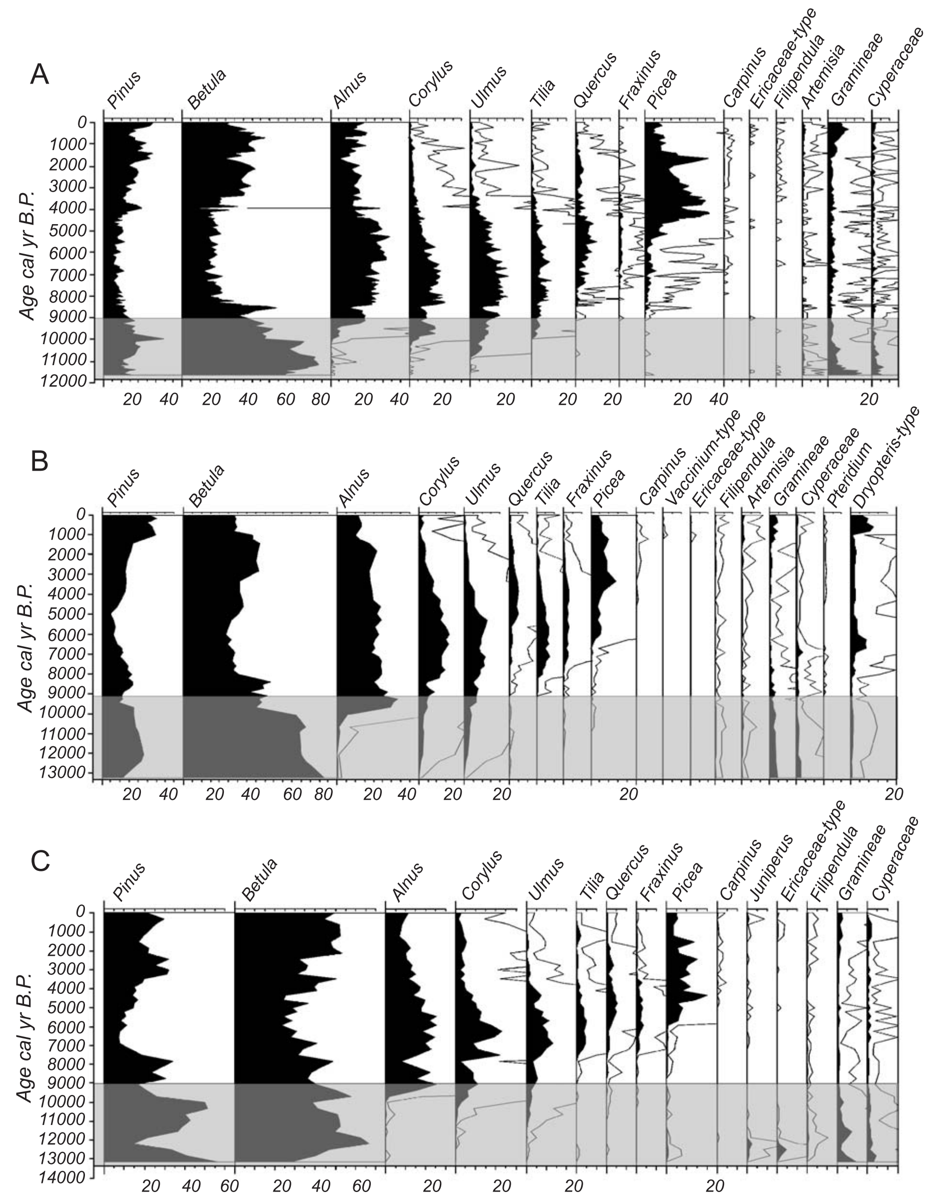

Fig. 3. Simplified pollen percentage diagrams from Lake Raigastvere (A), Lake Viitna (B), and Lake Ruila (C). Only the most common and important pollen and spore taxa are shown. Black silhouttes indicate the percentage values and the unshaded silhouettes 10× exaggerations. Pollen percentages are calculated from the total sum of terrestrial pollen and spore types. The grey shading indicates those parts of the diagrams where the chronological control is considered unreliable (see Fig. 2) and which are therefore not included in the T_{ann} reconstruction (Figs. 4 and 5).

decline steadily from 9000 cal yr B.P. (Berger, 1978; Crucifix et al., 2002). Consequently, a systematic analysis of the reconstructed Holocene climate changes requires identification of the relevant past circulation patterns. Currently, the variability of the seasonal features of the North Atlantic/European sector is strongly influenced by large-scale atmospheric circulation patterns (Marshall et al., 2001; Petterssen, 1949; Werner et al., 2000). In North Europe, the strong westerly circulation, often, but not always, associated with a high North Atlantic Oscillation index (Slonosky and Yiou, 2002), leads to positive temperature anomalies in winter and negative temperature anomalies in summer (Chen and Hellström, 1999; Hurrell and van Loon, 1997; Jacobeit et al., 2001; Marshall et al., 2001). In contrast, the strong meridional circulation, often associated with blocking anticyclonic conditions, leads to a more continental climate pattern and, depending on the location of the high-pressure cell, can result in high midsummer temperatures (Johannessen, 1970). On this basis, it is likely that the HTM was characterized by increasingly strong blocking anticyclonic atmospheric conditions over North Europe, and that the associated meridional atmospheric circulation pattern reached its maximum intensity at 6500–4000 cal yr B.P. with warm, dry summers.

Important, independent support for the inferred dynamics of the atmospheric circulation patterns on a continental scale

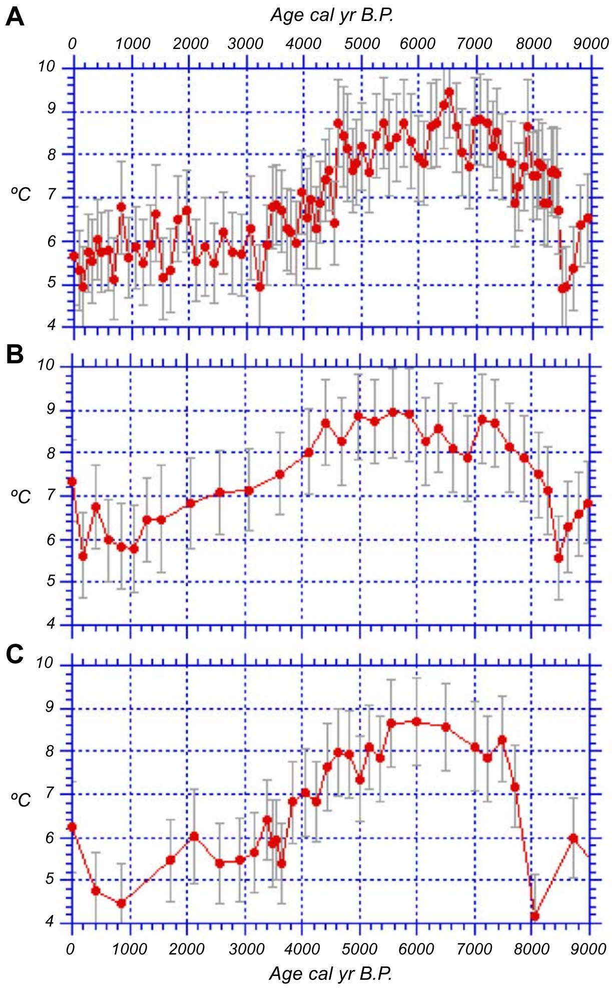

Fig. 4. Reconstructed annual mean temperatures during the past 9000 years. A = Lake Raigastvere, B = Lake Viitna, C = Lake Ruila. Sample-specific error bars are shown for each reconstructed value. Modern (1961–1990) observed annual mean temperature is 5.0°C at Lake Raigastvere and Lake Viitna and 5.5°C at Lake Ruila.

is provided by lake-level data (Harrison and Digerfeldt, 1993; Yu and Harrison, 1995). Lake-level changes indicate that at 10,000 cal yr B.P. atmospheric circulation over Estonia and much of North Europe was dominantly zonal, suggesting that the westerlies were both stronger and more northerly than they are today. During the mid-Holocene (ca. 7000–4000 cal yr B.P.), this pattern was replaced by blocking anticyclonic conditions with a more meridional flow pattern. This interpretation accords with ours. Yu and Harrison (1995) and Harrison et al. (1996) link the anticyclonic condition to the larger-than-present extent of the Baltic Sea at 8000–5000 cal yr B.P., but we regard it as more likely that the change toward anticyclonic conditions was caused by large-scale changes in latitudinal and longitudinal temperature and sea-level pressure gradients, by the North Atlantic sea-surface temperatures and oceanic circulation dynamics, by possible downstream influence of the melting of the Laurentide ice sheet on the North Atlantic oceanic and atmospheric circulation until ca. 8000 cal yr B.P., and by the synergistic impact of these changes, particularly on the intensity of the westerly air flow in North Europe.

Climate reconstructions: high-frequency events

The three reconstructions suggest that the most rapid changes of reconstructed T_{ann} took place during the early Holocene. All three reconstructions show a cooling of 1.5–2.0 °C which at Lake Raigastvere dates to 8600–8500 cal yr B.P., at Lake Viitna to 8400 cal yr B.P., and at Lake Ruila to 8100 cal yr B.P (Fig. 4). The colder reconstructed T_{ann} values are caused by similar pollen stratigraphical changes (Fig. 3), mostly by declines of *Corylus*, *Ulmus*, and *Alnus*, which suggest that the changes reflect a single, simultaneous event and that the apparent temporal offset may be an artefact caused by slightly erroneous radiocarbon dates and the resulting age–depth models. The rapid, transient cooling at 8600–8000 cal yr B.P. probably correlates with the distinct "8.2 ka event" widely recorded in the North Atlantic and northern Europe at 8400–8000 cal yr B.P. (Alley et al., 1997; Johnsen et al., 2001; Klitgaard-Kristensen et al., 1998; Nesje and Dahl, 2001; Tinner and Lotter, 2001). It was probably triggered by a ca. 400-year-long freshwater influx related to the collapse of the Laurentide ice sheet (Barber et al., 1999; Clark et al., 2001), subsequent reduction of North Atlantic deep water (NADW) formation, and weakening of the thermohaline circulation (THC) (Keigwin and Boyle, 2000; Renssen et al., 2001). However, at present there are no deep-sea or palaeochemical data showing that the production of the NADW was actually weaker at 8400–8000 cal yr B.P. (Clark et al., 2001; Keigwin and Boyle, 2000) and other possible causal mechanisms for the cooling have been proposed (Dean et al., 2002).

THC makes a major contribution to the heat budget of the North Atlantic region, raising annual mean surface

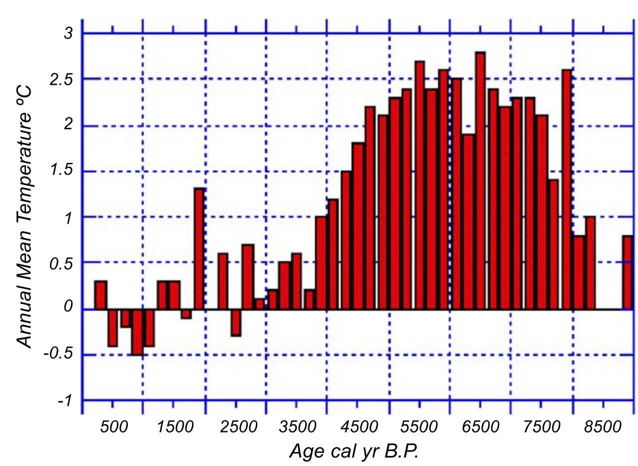

Fig. 5. Regionally averaged deviations of the reconstructed annual mean temperatures from the modern value, calculated as average values for 200-yr periods.

temperatures locally by up to 10°C (Manabe and Stouffer, 1995; Rahmstorf, 2000). It is still a matter of debate how far the warm anomaly created by the North Atlantic heat transport extends into the European continent, but it is likely that at least the northwestern European countries, including those in Fennoscandia, are warmed by several degrees, with the largest effect in winter (Rahmstorf, 2000). Provided that the cooling at 8400–8000 cal yr B.P. was due to freshwater influx from the Laurentide ice sheet and subsequent weakening of the THC, the 1.5–2.0°C cooling in Estonia at the same time provides evidence for a hemispheric-scale climatic teleconnection in Holocene climate history and, in agreement with model simulations (Renssen et al., 2001; Vellinga and Wood, 2002), demonstrates the sensitivity of North European climate to 25°E longitude to disturbances of the THC. However, more precise and better-dated records based on a number of independent proxies are needed to establish firmly the nature of climate change in Estonia at 8600–8000 cal yr B.P. and its links to the dynamics of the THC and atmospheric circulation changes.

Due to the location of Estonia on the border between the oceanic and continental climatic sectors of North Europe, its climate is particularly sensitive to expansion and intensification of the Siberian high-pressure cell and to the related winter outbreaks of polar air. Our results are thus of direct relevance for the current discussion about the millennial-scale changes of Eurasian (Siberian) high-pressure conditions. Meeker and Mayewski (2002) argue that the ion time-series data of the Greenland ice cores record major changes in marine (Icelandic low) and terrestrial (Siberian high) atmospheric circulation systems. Based on this, Rohling et al. (2002) argue that the approximate correspondence between the Holocene K+ record of the GISP2 ice core and SST record from the Aegean Sea shows evidence of consistently enhanced pressure in the winter/spring Siberian high at ca. 300 cal yr B.P., 2900 cal yr B.P., 5000 cal yr B.P., and 8300 cal yr B.P. Apart from the cooling at 8600–8000 cal yr B.P., there is no evidence of such patterns in our data or in other records from climatologically sensitive regions in North Europe (Heikkilä and Seppä, 2003; Seppä and Birks, 2001, 2002). This suggests that no episodic intensifications of the Siberian high pressure and related outbreaks of polar air have taken place in North Europe during the last 8000 yr.

Conclusions

The motivation for the current study is summarized as follows: (i) it provides the first quantitative climate reconstructions from Estonia, a country located in a zono-ecotone potentially sensitive to the climatic impacts of changes in the North-European atmospheric circulation patterns, (ii) the use of three reconstructions provides a more reliable basis to distinguish consistent patterns (signal) from random variability (noise) than one reconstruction, (iii) we can compare the present results with the high-resolution reconstructions from southern Finland (Heikkilä and Seppä, 2003) and the Fennoscandian treeline area (Seppä and Birks, 2001, 2002), carried out with the same methodology, and (iv) inclusion of a number of reconstructions makes it possible to carry out comparisons with climate model simulations and to present hypotheses about past atmospheric circulation patterns.

Our results show that the regionally averaged T_{ann} was close to the present values at 9000–8000 cal yr B.P. and reached its maximum level, ca. 2.5°C above the present, at 8000–4500 cal yr B.P. Since 4500 cal yr B.P. T_{ann} has gradually decreased to the present level. The current data indicate that the only major cold event during the past 10,000 years in Estonia was the transient cooling at 8600–8000 cal yr B.P. which may be correlated with the 8400–8000 cal yr B.P. cold event in the Greenland ice cores and in the deep-sea cores from the North Atlantic and tentatively linked with a weakening of the THC. No abrupt cold events can be detected during the past 8000 years. In general, the reconstructed Holocene temperature trends in Estonia are markedly similar with the earlier pollen-based T_{ann} reconstruction from southern Finland (Heikkilä and Seppä, 2003) as well as with the $\delta^{18}O$ record of the NorthGRIP ice core (Johnsen et al., 2001) and borehole temperature measurements of the GRIP ice core (Dahl-Jensen et al., 1998) in Greenland.

The predominant driver of the reconstructed climate patterns has arguably been solar insolation. However, there is an increasing body of evidence, based mostly on pollen data (Heikkilä and Seppä, 2003; Seppä and Birks, 2001, 2002), $\delta^{18}O$ records from calcareous lake sediments (Hammarlund et al., 2002, 2003; Shemesh et al., 2001), and lake-level reconstructions (Harrison and Digerfeldt, 1993; Harrison et al., 1996; Yu and Harrison, 1995), to suggest that the Holocene climate history of North Europe has been more complex with changes in T_{ann}, seasonality, and humidity that deviate from the monotonic astronomical solar forcing trend. As is the case with the annual- and decadal-scale climate variability in northern Europe during past centuries (Marshall et al., 2001; Slonosky et al., 2000; Werner et al., 2000), such a dynamism of the Holocene climate can be best explained by secular shifts in the dominance of past atmospheric circulation patterns over North Europe. Understanding the causal factors of these atmospheric circulation changes is a major challenge for the palaeoclimate and climate modeling community.

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