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# Spatial structure of the 8200 cal yr BP event in northern Europe

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**Abstract.** A synthesis of well-dated high-resolution pollen records suggests a spatial structure in the 8200 cal yr BP event in northern Europe. The temperate, thermophilous tree taxa, especially *Corylus*, *Ulmus*, and *Alnus*, decline abruptly between 8300 and 8000 cal yr BP at most sites located south of 61° N, whereas there is no clear change in pollen values at the sites located in the North-European tree-line region. Pollen-based quantitative temperature reconstructions and several other, independent palaeoclimate proxies, such as lacustrine oxygen-isotope records, reflect the same pattern, with no detectable cooling in the sub-arctic region. The observed patterns challenges the general view of the widespread occurrence of the 8200 cal yr BP event in the North Atlantic region. An alternative explanation is that the cooling during the 8200 cal yr BP event took place mostly during the winter and spring, and the ecosystems in the south responded sensitively to the cooling during the onset of the growing season. In contrast, in the sub-arctic area, where the vegetation was still dormant and lakes ice-covered, the cold event is not reflected in pollen-based or lake-sediment-based records.

## 1 Introduction

High-resolution records have revealed that abrupt climate changes were frequent during the last glacial when regional temperature changes of as much as 8°C to 16°C may have occurred in a decade or less (Severinghaus and Brook, 1998; Stocker, 2000; Alley et al., 2003; Schulz et al., 2004), but also during the early post-glacial period characterized by rapidly vanishing ice sheets (Clark et al., 2001, 2002). The precise origins and processes associated with these events remain controversial (Schulz et al., 2004), as do their spatial expressions (Alley et al., 2003; Wunsch, 2006), although most theories invoke the role of the Atlantic meridional overturning circulation (AMOC) and its sensitivity to freshening of the North Atlantic surface water by increased precipitation, runoff from surrounding landmasses, and abrupt fluxes of glacial melt-water (Clark et al., 2002).

Many records, especially from the North Atlantic region, provide evidence of a cold event at 8200 cal yr BP that represents a unique climatic feature within the last 10,000 years in terms of magnitude and abruptness (e.g. Alley et al., 1997; von Grafenstein et al., 1998; Klitgaard-Kristensen et al., 1998; Johnsen et al., 2001; Spurk et al., 2002; Veski et al., 2004; Alley and Àgústsdóttir, 2005). Although it has been suggested that the cooling was linked to reduced solar output (Rohling and Pälike, 2005), there is accumulating evidence that the primary cause of the cooling was a pulse a cold freshwater released by a sudden drainage of the proglacial Laurentide lakes in North America to the North Atlantic at about 8500 cal yr BP, leading to a transient freshening and cooling of the North Atlantic Surface Water (NADW). This probably resulted in a weaker AMOC, and a consequent reduction of the northward heat transport and associated heat release in the North Atlantic region. Compelling support for this hypothesis is provided by the record of Ellison et al. (2006), showing that the near-bottom flow speed of the Iceland-Scotland Overflow Water, an important component

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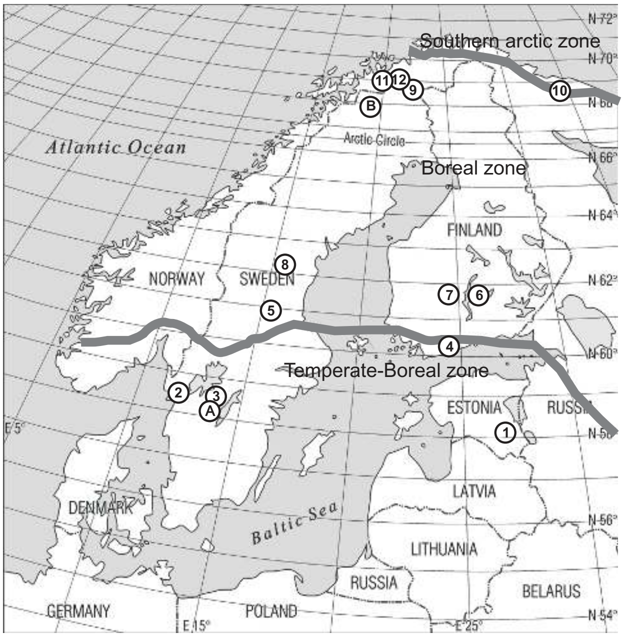


**Fig. 1.** The locations of sites (1–12) from which pollen-based temperature reconstructions used in this study have been obtained, together with sites (1, A, B) with oxygen-isotope records used for comparison (Table 1). The approximate boundaries of the main biomes in the region are shown. 1 = Lake Rouge, Estonia, 2 = Lake Trehörningen, Sweden, 3 = Lake Flarken, Sweden, 4 = Lake Arapisto, Finland, 5 = Lake Holtjärnen, Sweden, 6 = Lake Laihalampi, Finland, 7 = Lake Nautajärvi, 8 = Lake Klotjärnen, Sweden, Finland, 9 = Lake Tsuolbmajavri, Finland, 10 = Lake KP-2, Russia, 11 = Lake Dalmutladdo, Norway, 12 = Lake Toskaljavri, Finland, A = Lake Igelsjön, Sweden, B = Lake Tibetanus, Sweden.

of the AMOC, declined significantly at the onset of the cold event. This is the first firm palaeoceanographic evidence for a reduction of the NADW formation. This theory is supported by climate models which, in accordance with palaeoclimatic records, simulate maximum cooling in the North Atlantic region in response to the drainage of the Laurentide lakes (Alley and Àgústsdóttir, 2005; LeGrande et al., 2006; Wiersma and Renssen, 2006; Wiersma et al., 2006).

One of the key regions for investigating the continental-scale impacts of the 8200 cal yr BP cold event is Northern Europe, located downwind of the North Atlantic Ocean. The climatic conditions there are strongly dependent on the intensity of the North Atlantic Oscillation (NAO) and the associated westerly airflow, which is related to the strength of the AMOC and to the sea-surface temperatures in the North Atlantic (Hurrell, 1995; Rodwell et al., 1999; Hurrell et al., 2003; Wu and Rodwell, 2004; Stouffer et al., 20006). Thus, palaeoclimatic records from northern Europe can provide basis for testing potential processes associated with the 8200 cal yr BP cold event. Since the first reported occurrences of the cold event based, for example, on high-resolution pollen records (Snowball et al., 2001; Veski et al., 2004) and stable isotope records obtained from calcareous lake sediments (Hammarlund et al., 2003; Veski et al., 2004), the number of well-dated quantitative temperature reconstructions from this region has increased rapidly. This improving network of records with high time resolution now permits a more detailed spatial and temporal analysis of the climatic changes between 9500 to 7000 cal yr BP.

Here we examine high-resolution pollen-stratigraphical records produced with uniform methodology in northern Europe along a sector that ranges from 55 to 70° N latitude and from 18 to 26° E longitude and includes two significant climatic gradients: a primary south-to-north gradient of falling temperature and a secondary west-to-east gradient of decreasing precipitation and oceanicity. We particularly aim to investigate the spatial patterns of vegetational and climatic change in order to observe if the evidence for the cold event shows consistent features or whether there are geographical differences in the amplitude or occurrence of the event along the two climatic gradients. In addition, we compare the high-resolution pollen records with other, independent palaeoclimatic records of comparable time resolution, in particular oxygen-isotope records obtained on lacustrine carbonates.

## 2 Material and methods

There are numerous pollen diagrams available from northern Europe, but only a fraction of them have sufficient temporal resolution and chronological control to allow assessment of climate events of 200–300 years duration. Pollen records from 12 sites with adequate resolution and reliable chronologies were selected for this study (Fig. 1, Table 1). The number of analysed pollen samples for the last 10 000 years ranges from 57 (Lake Trehörningen) to over two hundred (Lake Nautajärvi 260 and Lake Rouge 237), and the records selected have a minimum of six radiocarbon dates and in general smooth age-depth models. Two records (Lakes Nautajärvi and Rouge), have exceptionally precise chronologies based on annually laminated sediments (Ojala et al., 2003; Veski et al., 2004). At all sites the pollen percentage values have been calculated on the basis of the total sum of all terrestrial pollen and spore types.

In addition to the examination of individual pollen curves, pollen-stratigraphical data were used to derive a quantitative temperature record for each site for the early- to mid-Holocene. Two different temperature parameters were used in the quantitative reconstructions. July mean temperature ($T_{\mathrm{jul}}$) was reconstructed at four sites located in the northern tree-line region, whereas annual mean temperature ($T_{\mathrm{ann}}$) was estimated for the rest of the sites located in the central or southern parts of the study area. The reason for this is that in the far north the growing season is confined to three or four summer months (MJJA) and a vegetation-based proxy such as pollen arguably predominantly represents summer

**Table 1.** Geographical locations and references of the pollen and oxygen-isotope records discussed in the study.

| | Lat. | Long. | Calibration model | Reference |
|---|---|---|---|---|
| 1 Rouge, Estonia | 57°44′ | 26°54′ | $T_{ann}$-3 | Veski et al., 2004 |
| A Igelsjön, Sweden | 58°28′ | 13°44′ | | Hammarlund et al., 2003, 2005 |
| 2 Trehörningen, Sweden | 58°33′ | 11°36′ | $T_{ann}$-3 | Antonsson and Seppä, 2007 |
| 3 Flarken , Sweden | 58°33′ | 13°44′ | $T_{ann}-3$ | Seppä et al., 2005 |
| 4 Arapisto, Finland | 60°35′ | 24°05′ | $T_{ann}-2$ | Sarmaja-Korjonen and Seppä, 2007 |
| 5 Holtjärnen, Sweden | 60°39′ | 15°56′ | $T_{ann}-3$ | this paper; Giesecke, 2005 |
| 6 Laihalampi, Finland | 61°29′ | 26°04′ | $T_{ann}-1$ | Heikkilä and Seppä, 2003 |
| 7 Nautajärvi, Finland | 61°48′ | 24°41′ | $T_{ann}-3$ | Ojala et al., 2007[1] |
| 8 Klotjärnen, Sweden | 61°49′ | 16°32′ | $T_{ann}-3$ | this paper; Giesecke, 2005 |
| B Tibetanus, Sweden | 68°20′ | 18°42′ | | Hammarlund et al., 2002 |
| 9 Tsuolbmajavri, Finland | 68°41′ | 22°05′ | $T_{jul}$ | Seppä and Birks, 2001 |
| 10 KP Lake, Russia | 68°48′ | 35°19′ | $T_{jul}$ | this paper; Gervais et al., 2002 |
| 11 Dalmutladdo, Norway | 69°10′ | 20°43′ | $T_{jul}$ | Bjune et al., 2004 |
| 12 Toskaljavri, Finland | 69°12′ | 21°28′ | $T_{jul}$ | Seppä and Birks, 2002 |

[1] Ojala, A. E. K., Alenius, T., and Seppä, H.: Integration of the clastic-organic varve record from Finland with a pollen-based climate reconstruction for solving the Holocene seasonal temperature patterns in the high latitudes, The Holocene, submitted, 2007.

temperature conditions. No such generalization can be made in more southern parts of Fennoscandia, however, because there the growing season is considerably longer, starting often in March or April and continuing to October (Rötzer and Chmielewski, 2001; 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). Thus the pollen records represent a mixture of taxa with different temperature requirements in relation to the seasons and annual mean temperature is probably a better justified climatic parameter to be reconstructed from pollen data in southern and central Fennoscandia than July or summer (JJA) mean temperature (Seppä et al., 2004).

Pollen-based temperature reconstructions were based on transfer functions derived from North-European pollen-climate calibration sets. These data sets are based on modern climate data from the Climate Normals period 1961–1990 and on modern pollen samples, collected from top surface sediment samples of small to medium-sized lakes. All samples were selected, collected, and analysed with standardised methods (see Seppä et al., 2004). The $T_{jul}$ records were based on 164 surface samples from Norway, 27 samples from northern Sweden, and 113 samples from Finland (Seppä and Birks, 2001). The $T_{ann}$ reconstructions were carried out using three different calibration subsets. Subset 1 consists of 113 samples from Finland, subset 2 includes 24 samples from Estonia in addition to the 113 Finnish samples (Seppä et al., 2004), and subset 3 includes 37 samples from Sweden in addition to the 113 Finnish samples and the 24 Estonian samples. Table 1 indicates which subset was used for each fossil record. The transfer functions were developed using weighted averaging partial least squares (WA-PLS) regression, a non-linear, unimodal regression and calibration technique commonly used in quantitative environmental reconstructions (ter Braak and Juggins, 1993; Birks, 1995, 2003). WA-PLS was implemented by the program CALIBRATE (S Juggins and CJF ter Braak unpublished program). All terrestrial pollen and spore types were included in the models, and were square-root transformed to stabilize their variances and to maximize “the signal to noise” ratio (Prentice, 1980). The performance statistics of the pollen-climate calibration sets are given in Table 2. In general the statistics indicate high performance of the models relative to other corresponding models based on various biological proxy techniques (Birks and Seppä, 2004).

**Table 2.** The performance statistics of the pollen-climate calibration models used for producing the temperature records. RMSEP = root mean square error of prediction, $R^2$ = coefficient of determination between the observed temperature and that predicted by the model.

| Model | number of samples | RMSEP | $R^2$ | max. bias |
|---|---|---|---|---|
| $T_{jul}$ | 304 | 0.99°C | 0.71 | 3.94°C |
| $T_{ann}-1$ | 113 | 0.91°C | 0.85 | 2.12°C |
| $T_{ann}-2$ | 137 | 0.89°C | 0.88 | 2.13°C |
| $T_{ann}-3$ | 174 | 0.95°C | 0.88 | 2.10°C |

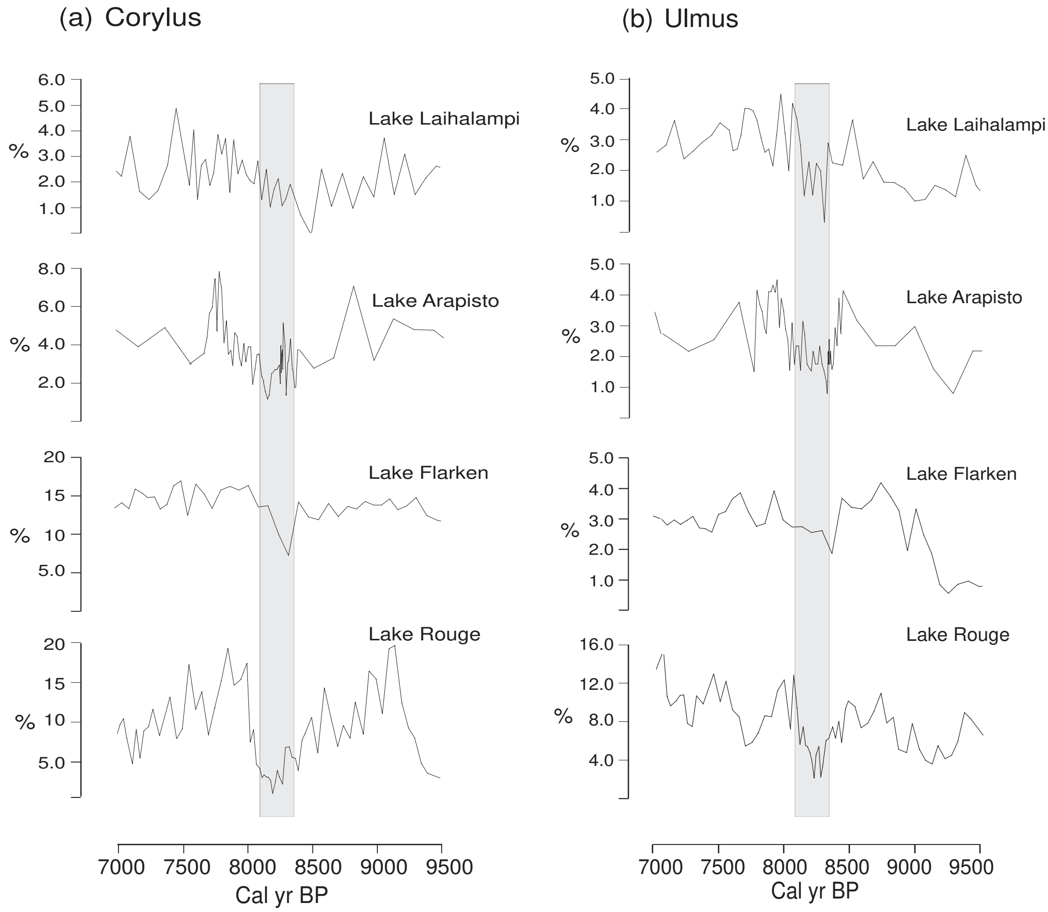


**Fig. 2.** *Corylus* and *Ulmus* pollen percentage curves at 9500 to 7000 cal yr BP from four sites in the southern and central parts of the study area. No records of these pollen types are shown from the sites in the far north of Europe because these thermophilous taxa do not occur there. See Fig. 1 for the locations of the sites.

## 3 Results

To show the details of the pollen-stratigraphical changes during the 8200 cal yr BP event in northern Europe, we focus on key pollen types. In northern Europe pollen types that mostly decline during the event, indicating either reduced populations and/or pollen productivity, likely caused by the cooling, are the thermophilous deciduous tree taxa, predominantly *Corylus*, *Ulmus* and *Alnus*. The types that usually increase are *Betula* and/or *Pinus* (Snowball et al., 2001; Veski et al., 2004; Seppä et al., 2005; Sarmaja-Korjonen and Seppä, 2007). The percentage pollen curves of *Corylus* and *Ulmus* are shown in Fig. 2. As these temperate deciduous tree taxa do not occur in the pollen records from the tree-line region, curves are shown only for the sites that are located in the southern and central part of the research area. Both taxa show a distinct and abrupt decline during the event at four sites, all located south of 61° N. *Corylus* especially seems to respond strongly to the sudden cooling, its values dropping at Lake Rouge, for example, from 10–15% to below 5% at 8250–8050 cal yr BP, and the decline is almost equally distinct at the three other southernmost sites. The decline of *Ulmus* is of similar magnitude apart from the weak signal at Lake Flarken. At Lakes Laihalampi and Nautajärvi, both located north of 61° N, *Corylus* does not decline during the cold event, but there is a relatively clear decline of *Ulmus* at Lake Laihalampi (Fig. 2).

*Alnus* pollen percentage curves are shown from eight sites (Fig. 3). *Alnus* is selected because it is the only thermophilous deciduous tree taxon whose pollen values continuously exceed 1% at all sites, including those at the northern tree-line. *Alnus* pollen is produced by two tree species, *A. incana* and *A. glutinosa*, of which the former is found up to 68° N. The comparison of *Alnus* pollen records indicates a south to north gradient. There is a clear decline of *Alnus* values at the southern sites, especially at Lakes Rouge and Arapisto, whereas records from central Fennoscandia indicate little or no decline. *Alnus* pollen records from the Arctic tree-line region show no evidence of a decline.

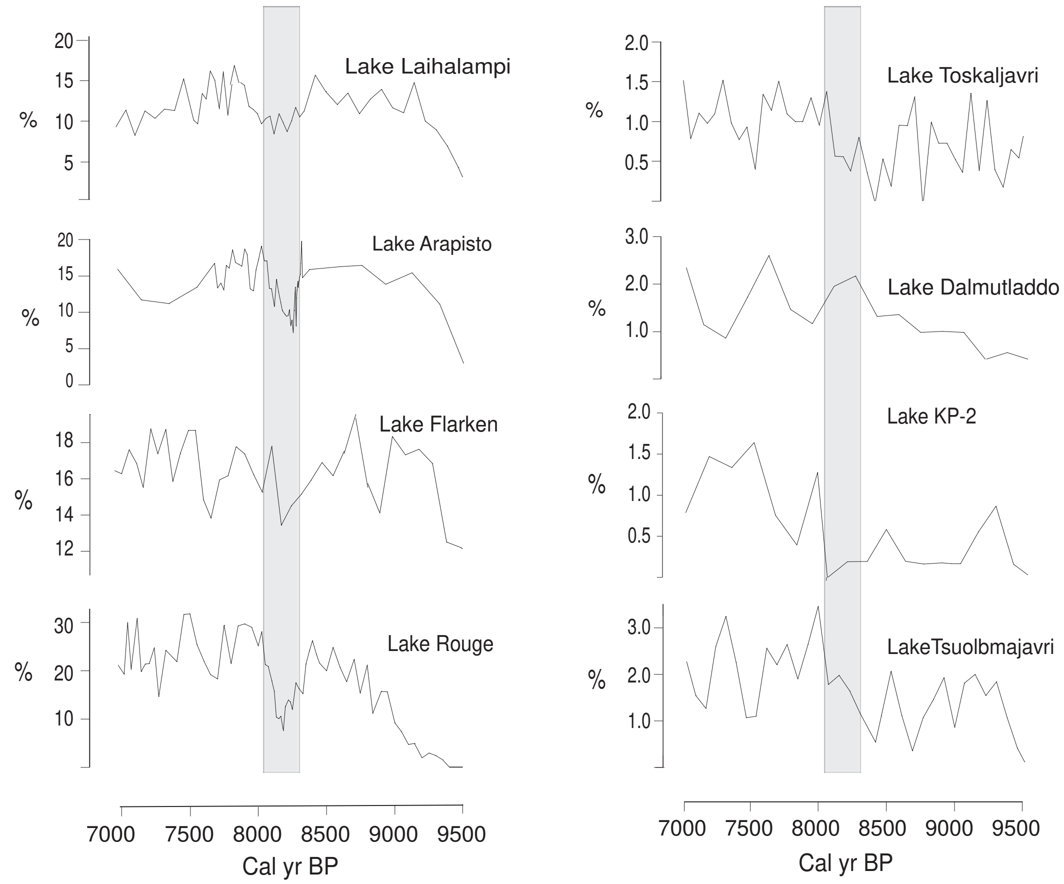


**Fig. 3.** *Alnus* pollen percentage curves at 9500 to 7000 cal yr BP from the eight sites in northern Europe. Sites on the left, showing a decline of the *Alnus* values at about 8300–8100 cal yr BP, are from the southern part of the study area, whereas sites of the right, with no clear decline of *Alnus*, are from the North-European tree-line region. See Fig. 1 for the locations of the sites.

The pollen-based quantitative $T_{ann}$ reconstructions for the period 9500–7000 cal yr BP show broadly the same pattern as the key indicator pollen types (Fig. 4). The results indicate a consistent cooling centred around 8200 cal yr BP at the southernmost sites, especially at Lake Rouge in Estonia, Lake Flarken in Sweden, and Lake Arapisto in Finland. At all these sites there is a temperature drop of 0.5 to 1.5°C. The cooling begins abruptly at about 8300 cal yr BP, lasts 200 to 300 years, and ends with a sudden temperature rise at about 8000 cal yr BP. As the age-depth models for most sites were based on calibrated radiocarbon dates, it is realistic to relate the slight temporal differences between the records to the inevitable imprecision of the chronologies. In contrast, the $T_{ann}$ reconstructions from Lakes Holtjärnen, Laihalampi and Nautajärvi, all located north of 61° N, show weak or no evidence of cooling at 8300–8000 cal yr BP. $T_{jul}$ records from the Fennoscandian tree-line region do not provide any evidence for a temperature change during the event. All four sites indicate a steady rise of $T_{jul}$ from 9400 cal yr BP toward the mid-Holocene, with variability that is not consistent between the records. Hence, the quantitative reconstructions follow the same spatial pattern as the key pollen types, namely that there is a distinct cooling between 8300 and 8000 cal yr BP in southern Fennoscandia and in the Baltic countries, weak or no cooling in central Fennoscandia, and no evidence of cooling in the tree-line region of northern Europe (Figs. 5 and 6).

A synthesis of the pollen-based evidence shows strikingly that the four records where the cooling is clearest are from the southernmost sites, located south of 61° N. Biogeographically, they are located in the temperate-boreal (boreonemoral) zone today (Fig. 1) where the vegetation is characterized by the occurrence of nemoral thermophilous tree species such as *Tilia cordata*, *Quercus robur*, *Corylus avellana*, *Ulmus glabra*, *Acer platanoides*, and *Fraxinus excelsior*. All these tree species reach their northern distribution limits close to the border of the nemoral and boreal vegetation zones. In contrast, all the records where there is no cooling at 8200 cal yr BP are from sites that are located north of 61° N. Lakes Laihalampi and Nautajärvi are located in the southern boreal zone, where the dominant forest type is a mixture of conifers and birch, and Lakes Tsuolbmajavri, KP-

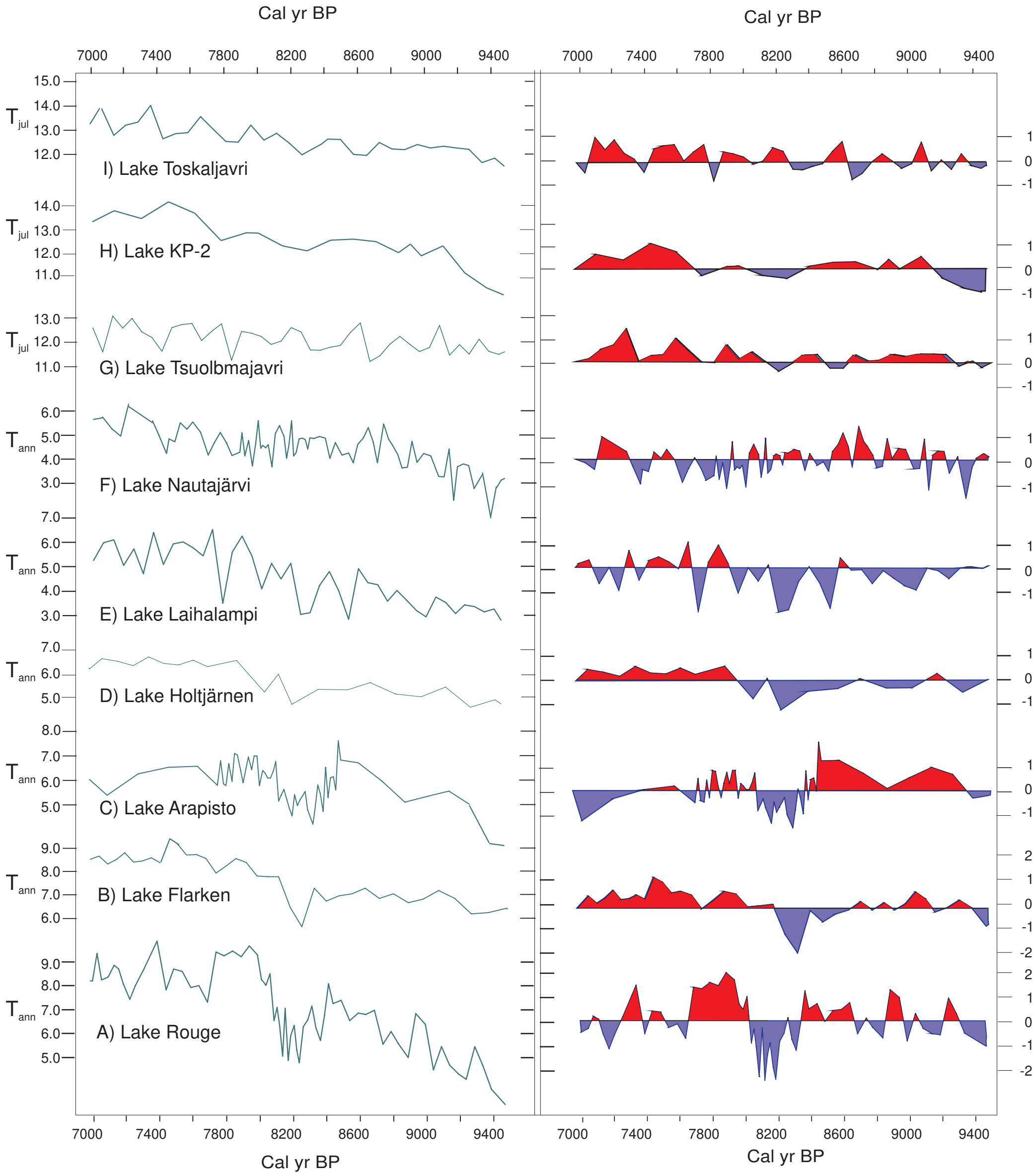


**Fig. 4.** Pollen-based $T_{\mathrm{ann}}$ and $T_{\mathrm{jul}}$ reconstructions at nine sites for the time period 9500 to 7000 cal yr BP arranged from the southernmost to the northernmost site. Original records are shown on the left hand side. The panel on the right shows the residuals after detrending the records with a third-order polynomial curve. See Figure 1 for the locations of the sites.

2, Dalmutladdo, and Toskaljavri are all situated within the northern ecotone of the boreal zone, with significant arctic-alpine components in the surrounding vegetation.

## 4 Discussion

The relatively mild winter climate of northern Europe is a result of the influence of the Atlantic Ocean and the westerly airflow over the continent (Seager et al., 2002; Sutton et al., 2003). This influence is particularly strong during the positive phase of the NAO with strong westerly wind, greater advection of moist air off the Atlantic and onto the continent, and high temperature and precipitation even in northernmost Fennoscandia (Hurrell et al., 2003; Kryjov, 2004; Cook et al., 2005; Jaagus, 2006). The reverse holds true when the NAO has a negative mode and the westerly airflow is replaced by anticyclonic conditions and a predominantly continental airflow. The most probable scenario for cooling during the 8200 cal yr BP cold event in northern Europe is that the sudden flux of cold freshwater perturbed the AMOC and led to lower sea-surface temperatures in the North Atlantic. As a consequence the heat transport from the Atlantic onto

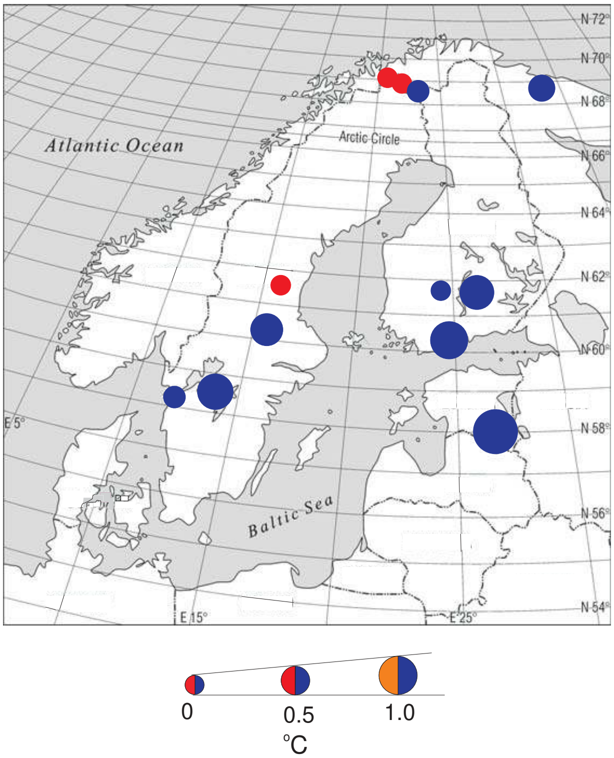


**Fig. 5.** Map showing the geographical pattern of the temperature change. The blue colour indicates cooling and red colour warming during the 8200 cal yr BP event. The temperature deviation at the event is calculated as the difference between the mean temperature at 8350 – 8050 cal yr BP and the mean for the periods 8850–8350 and 8050–7550 cal yr BP.

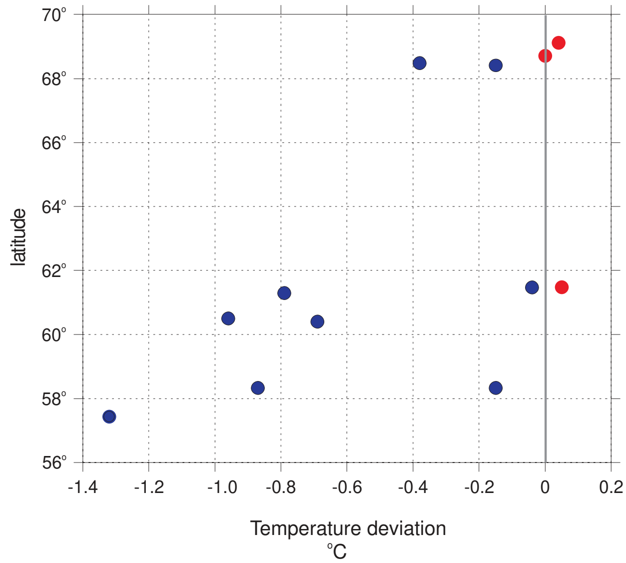


**Fig. 6.** Temperature deviation during the 8200 cal yr BP event along a south-to-north gradient. Data from all 12 pollen-based records are included. The deviation is calculated by comparing the reconstructed average July mean temperature or annual mean temperature at 8350–8050 cal yr BP relative to the mean temperature of the periods 8850–8350 cal yr BP and 8050–7550 cal yr BP.

the continent decreased. The reason for this may have been a weaker generation of migratory cyclones and, in general, a weaker flow of mild oceanic air over northern Europe during winter (Veski et al., 2004; Hammarlund et al., 2005; Seppä et al., 2005). A related hypothesis suggests that the weaker oceanic airflow may have resulted from a major expansion of sea-ice cover in the North Atlantic and an associated reduction of the advection of heat from the ocean to the atmosphere (Wiersma and Renssen, 2006).

Against this background, it can be argued that the reconstructed longitudinal gradient towards a weaker signal in the North reflects realistically a geographical pattern of the magnitude of the 8200 cal yr BP event. The cold event may have been caused by weakened westerly circulation during winter, so that the resulting decrease in oceanicity was particularly influential south of ~60° N latitude and less significant in the far north of Fennoscandia. This explanation may seem paradoxical, given the presently strong influence of the North Atlantic and the NAO on the climate of northern Fennoscandia, particularly on the eastern side of the Scandes Mountains (Hurrell et al., 2003). It is also inconsistent with the results of modelling studies focusing on the 8200 cal yr BP event (Renssen et al., 2001, 2002; Alley and Àgústsdóttir, 2005; LeGrande et al., 2006; Wiersma and Renssen, 2006), all indicating major, wide-spread cooling in the North Atlantic and the eastern Atlantic seaboard in response to weakening of the AMOC. Interestingly, however, the multimodel ensemble simulation based on models ranging from the earth system models of intermediate complexity to fully coupled atmosphere-ocean general circulation models indicates that a moderately small freshwater flux of 0.1 Sv (1 Sv=$10^6$ $m^3$ $s^{-1}$) may lead to a strong cooling in the North Atlantic south of Greenland but a 1–2°C warming over the Barents Sea and the Nordic Sea east of Svalbard and no significant temperature change on the northern Fennoscandian mainland (Stouffer et al., 2006).

An alternative hypothesis can also explain the observed geographical pattern in the records. Model simulations consistently indicate that the 8200 cal yr BP event was predominantly a winter (DJF) and spring (MAM) event, as is also supported by the majority of palaeoclimate records from various parts of Europe (Alley and Àgústsdóttir, 2005; Wiersma and Renssen, 2006). The tree taxa that show strong responses to the 8200 cal yr BP event, *Alnus*, *Corylus*, and *Ulmus,* start flowering in early spring, in central Europe often in February–March and in southern and central Fennoscandia in March–April (Jäger et al., 1996; Kasprzyk et al., 2004). The start of their flowering and plant development is in general dependent on air temperature (Wielgolaski, 1999; Aasa et al., 2004) and if an abrupt change to cold winters and cold

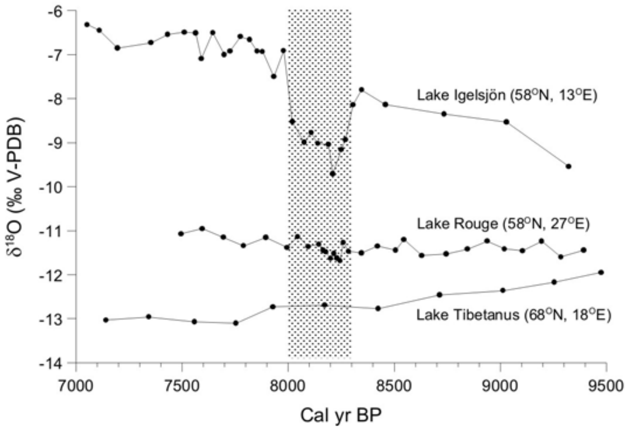


**Fig. 7.** Comparison of $\delta^{18}$O records obtained on fine-grained sedimentary calcite from small lakes (0.5–4 ha) in the study area, Lakes Igelsjön (Hammarlund et al., 2003, 2005; Seppä et al., 2005), Rouge (Veski et al., 2004), and Tibetanus (Hammarlund et al., 2002). See Fig. 1 for the locations of the sites.

early springs with frequent frosts took place at the beginning of the event, it may have led to major damage to flowers and male catkins, resulting in reduced pollen productivity, sexual regeneration, and population sizes. This hypothesis is supported by investigations of the relationships between modern phenological phenomema and climate patterns (Kramer et al., 2000) and by modern pollen monitoring studies in Europe. For example, cold weather conditions during winter and spring result in a reduction of the annual pollen productivity and a delay of the start of the pollen season of tree species that are favoured by high temperatures, such as *Alnus*, *Corylus* and *Ulmus* (Andersen, 1972; Frenguelli, 1993; Jäger et al., 1996; Frei, 1998; Spieksma et al., 2003; Kasprzyk et al., 2004). The clear decline of the pollen percentages of these taxa may therefore reflect their phenological inability to adapt to an abrupt lowering of the winter and early spring temperatures during the 8200 cal yr BP event.

In contrast, northernmost Fennoscandia, north of 68° N, is characterized by a markedly different climatic and phenological situation. Here winter conditions with subzero diurnal mean temperatures and abundant snow remain long into April. The growing season starts in late May or early June (Linderholm et al., 2006) and its duration is typically only 100–150 days. Winter temperatures can be extremely low, down to below $-40$°C, and only the most frost-resistant tree species thrive, i.e. the main northern forest components *Pinus sylvestris, Picea abies*, and *Betula (B. pubescens* and *B. pendula*) (Grace et al., 2002). For example, climate chamber experiments have shown that *Pinus sylvestris* can tolerate temperatures as low as $<-72$°C during the winter while its needles can be lethally damaged when exposed to $-10$°C during the summer (Repo, 1992; Beck et al., 2004). The continental conifer species *Picea abies* is also adapted to cold winters as its radial growth is positively correlated with low winter temperatures and negatively correlated with high winter temperatures (Mäkinen et al., 2000). Of the deciduous tree taxa that show strong responses to the 8200 cal yr BP event in the southern part of the study region, only *Alnus* grows in northern Fennoscandia, albeit not north of about 68° N. However, *Alnus incana* is less sensitive to cold periods during the late winter and early spring in the far north, because of phenological adaptation of the beginning of flowering, bud break, and leaf expansion, these processes taking place usually in May to June, about two months later than in the south.

If the cooling took place mostly during winter and early spring, the occurrence, regeneration, and pollen productivity of the northern-boreal tree taxa may not have been significantly affected, and the lack of pollen-stratigraphical responses to the cold event may reflect largely unaltered climatic conditions during the short growing season of these taxa. Many other proxy records that are independent of vegetation patterns indicate a weaker signal for the 8200 cal yr BP event towards northern Fennoscandia. In western Scandinavia records based on loss-on-ignition analysis of lake sediments reveal a substantial climatic perturbation and clearly decreased aquatic productivity at about 8200 cal yr BP (Nesje et al., 2001; Bergman et al., 2005). However, in northern Fennoscandia the same methods do not display equally clear evidence for a cooling. For example, a high-resolution reconstruction of Holocene equilibrium-line altitude changes from the Lyngen peninsula on the northwestern coast of Norway does not show any sign of glacier growth between 8500 and 7500 cal yr BP (Bakke et al., 2005), neither does the record of bacterial magnetite from the Lyngen peninsula indicate any cold excursion during the 8200 cal yr BP event (Paasche et al., 2004). Chironomid-based temperature reconstructions from northern Finland and Sweden do not indicate any cooling between 8300 and 8000 cal yr BP (Rosén et al., 2001; Bigler et al., 2002, 2003; Korhola et al., 2002; Seppä et al., 2002; Larocque and Hall, 2003, 2004). Similarly, quantitative and qualitative reconstructions from the Kola Peninsula in northwestern Russia do not document any cooling at 8200 cal yr BP (Jones et al., 2004). Korhola et al. (1999) and Bigler et al. (2006) report a cooler period at about 8300 to 8000 cal yr BP in diatom-based $T_{\mathrm{jul}}$ reconstructions from the northwestern Fennoscandian tree-line region, but it is unclear whether these results represent summer cooling or are more related to changes in the length of the lake ice-cover season, which may be the most important climate-related feature reflected in diatom records from alpine regions (Lotter and Bigler, 2000; Sorvari et al., 2002).

Independent evidence of a considerable influence of winter cooling on the general climatic character and expression of the 8200 cal yr BP event in parts of northern Europe is provided by oxygen-isotope records ($\delta^{18}$O) obtained on fine-grained sedimentary calcite from small lakes in the study area. Although relatively few records are available, and in spite of complications arising from site-specific hydrological characteristics, some relevant conclusions can be derived from such a comparison (Fig. 7). The Lake Igelsjön $\delta^{18}$O record from southern Sweden (Fig. 1) is sensitive to changing hydrology, with periods of $^{18}$O-enrichment reflecting mainly elevated evaporation/inflow ratio of the basin under warm and dry summer conditions (Hammarlund et al., 2003). The strong isotopic response to the 8200 cal yr BP event therefore predominantly reflects an increase in net precipitation during the summer (Hammarlund et al., 2005), perhaps augmented by a generally shorter ice-free season. However, as demonstrated by Seppä et al. (2005), about 40% of the decrease in $\delta^{18}$O during the event can be attributed to a depletion in $^{18}$O of annual precipitation and groundwater, which is likely coupled to a large extent to colder and longer winters. The effect of such a change in winter conditions is also manifested as a slight depletion in $^{18}$O at 8200 cal yr BP in the Lake Rouge record from southern Estonia (Veski et al., 2004), which more directly reflects $\delta^{18}$O of precipitation.

Oxygen-isotope data of comparable resolution are not available from northern Fennoscandia, but the Lake Tibetanus $\delta^{18}$O record from northernmost Sweden, close to the Atlantic coast (Hammarlund et al., 2002), does not reflect any climatic change at this stage. Supportive evidence of the absence of an oxygen-isotope response at 8200 cal yr BP in this part of Fennoscandia is provided by the more highly resolved $\delta^{18}$O record from the SG93 speleothem near the Arctic Circle in Norway (Lauritzen et al., 1999). Although interpreted differently by the authors, the SG93 record exhibits a general depletion in $^{18}$O with time during the early Holocene, consistent with the long-term evolution of $\delta^{18}$O of precipitation as inferred from the Lake Tibetanus data (Hammarlund et al., 2002). It cannot be excluded that the potential effect of a pronounced winter cooling, and an associated depletion in $\delta^{18}$O of precipitation, on the two latter records from northwestern Scandinavia was offset by lowered summer temperatures during the event, leading to $^{18}$O-enrichment of lacustrine and speleothem calcite. However, it appears likely from these data that the weakening of the AMOC during the 8200 cal yr BP event and the associated cooling of the North Atlantic Ocean, induced a southward displacement of the Polar Front and the westerlies in winter (Magny et al., 2003), giving rise to a general anti-cyclonic circulation pattern over northern Europe. Such a scenario may be invoked to explain the greater inferred cooling in the southern part of the study area in response to enhanced continentallity (Fig. 5), as well as parts of the stronger isotopic response in southern Sweden as compared to Estonia (Fig. 7), where the change was of lesser magnitude due to a relatively continental baseline climate.

## 5 Conclusions

The assessment of the regional impact of the 8200 cal yr BP event is based on a survey of records of positive evidence in the increasingly dense network of temperature-sensitive proxy records. Such a survey is a delicate and difficult process, as the palaeorecords are typically noisy and it is difficult to distinguish regionally restricted and representative anomalies. Furthermore, ignoring records with negative evidence can cause bias in such an assessment. In northern Europe, a synthesis of negative and positive evidence from pollen-based temperature reconstructions indicates a spatial pattern in the 8200 cal yr BP event, with more distinct evidence of the cooling in the Baltic countries and in southern Fennoscandia than in the central and northernmost parts of Fennoscandia and adjacent areas. Given the evidence of the wide-spread nature of the 8200 cal yr BP event in the North-Atlantic region (Alley and Àgústsdóttir, 2005; Wiersma and Renssen, 2006) and the Barents Sea (Duplessy et al., 2004), a cooling probably took place all over northern Europe, including the tree-line region, but, as the cooling was predominantly a winter and spring event, taking place before the start of the growing season or before the break-up of lake ice in the north, it is not recorded in the quantitative and qualitative climate records obtained from the far north of northern Europe. However, on the basis of evidence presented here we cannot rule out a latitudinal gradient in the magnitude of the event, with a more pronounced cooling in the south and less or no cooling in the north. It may be possible to test these two hypotheses in the future by developing and applying specific palaeoecological techniques such as analyses of sedimentary chrysophyte cysts (Kamenik and Schmidt, 2005; Pla and Catalan, 2005), clacoceran ephippia (Sarmaja-Korjonen, 2004), or diatom records from alpine lakes (Lotter and Bigler, 2000) as they may have the potential for reflecting the length of the winter ice-cover of the lakes, hence providing insights into the winter temperature changes during the event.

If the muted response to the 8200 cal yr BP event at the tree-line sites results from the insensitivity of the palaeoclimatic records, then an important implication is that the northern tree-line regions, and in a more general sense, the cold regions of the Earth, may not always be optimal targets for palaeoclimatic reconstruction. In these regions the biological activity is to a great extent restricted to the short growing season during the summer months, whereas during the long and cold winter both terrestrial and aquatic ecosystems are dormant and therefore less sensitive to temperature changes. Consequently, biological proxy techniques such as pollen and chironomid records reflect predominantly summer temperatures in these regions. The same proxies in regions with a longer biologically active period, such as the ecotone between the temperate and boreal zones, are more sensitive to spring, autumn and perhaps winter temperatures. The evidence associated with the 8200 cal yr BP event provides therefore a prime example of the importance of site selection in palaeoclimatological and palaeoecological studies.

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