Holocene Climate Reconstructions from the Fennoscandian Tree-Line Area Based on Pollen Data from Toskaljavri

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Due to its location on the eastern North Atlantic seaboard, the Fennoscandian tree-line area is an ideal area to use biological proxies to assess the relative roles of the Scandinavian ice sheet and of oceanic, atmospheric, and astronomic forcings on regional climate history. Here we report pollen-based July mean temperature ($T_{jul}$) and annual precipitation ($P_{ann}$) reconstructions from a sediment core from a high-altitude tree-line lake in northwestern Finland. The reconstructions suggest that at 9600–8300 cal yr B.P. $T_{jul}$ values were low but steadily rising while $P_{ann}$ was high. The period of warmest summers, with $T_{jul}$ values ca. 1.8–1.6°C higher than at present, occurred at ca. 8000–6500 cal yr B.P. Since then climate has become gradually cooler. $T_{jul}$ values during the “Medieval Warm Period” (ca. 1400–1000 cal yr B.P.) were ca. 0.8°C higher than at present but decreased rapidly to the low “Little Ice Age” levels at 80 cal yr B.P. We compare these results with an earlier pollen-based climate reconstruction from the same region. The reconstructions indicate a similar general Holocene $T_{jul}$ pattern with lower values in the reconstruction from the high-altitude lake. However, most of the small-scale variations are not synchronous, suggesting that they may represent noise rather than signal in our data.

Key Words: Holocene; pollen; temperature; precipitation.

INTRODUCTION

The climate of northern Fennoscandia is anomalously mild in relation to its latitudinal location. At 70°N latitude the annual incoming solar radiation at the Earth’s surface is only ca. 70,000 cal/cm², whereas the outgoing radiation is ca. 136,000 cal/cm² (Wallén, 1970). In northern Europe, the deficit is partly balanced by oceanic and atmospheric circulation processes that carry heat from lower latitudes. The heat transport is driven by the North Atlantic thermohaline circulation, an oceanic system carrying warm surface waters from the subtropical Atlantic to the North Atlantic (Manabe and Stouffer, 1999). From there, the heat is transported to continental Europe by westerly air flow, especially as migratory cyclones (Lamb and Johnson, 1959; Johannessen, 1970; Kozuchowski, 1993). Thus, the climate and the associated location of the tree line in northern Fennoscandia at roughly 70°N are critically dependent on the intensity and dynamics of the North Atlantic ocean–atmosphere circulation.

Recently, much paleoclimatic interest has been directed to the short- and long-term dynamics of North Atlantic atmospheric circulation patterns (Hurrell, 1995). The most conspicuous demonstration of this dynamism is the North Atlantic Oscillation (NAO), an oscillatory atmospheric air pressure pattern, in which a large-scale air displacement takes place between the Azores subtropical high and the low pressure system of Iceland in the North Atlantic (Stephenson et al., 2000). During a high-index mode the air pressure difference between the systems is enhanced while the systems exhibit a synchronous shift northward (Paeth et al., 1999). Over the Holocene time scale, an even larger scale dynamism in the North Atlantic has been suggested, namely cyclic weakenings in the North Atlantic thermohaline circulation and an associated occurrence of short-term cold periods (Bond et al., 1997).

Here we present a high-resolution Holocene climate reconstruction from Toskaljavri, a tree-line lake situated in the continental sector of northern Fennoscandia, but only ca. 70 km from the Norwegian Sea coast (Fig. 1). We use a recently developed pollen-climate reconstruction model and a new pollen stratigraphy to derive quantitative estimates of annual precipitation and July mean temperature. We used a similar approach to derive July mean temperature and annual precipitation reconstructions from Tsuolbmajavri, another small lake located 60 km south-east of Toskaljavri, in the same climatic and vegetational area, but at a slightly lower altitude (Seppä and Birks, 2001). The aim of this paper is to evaluate the climatic interpretation based on the Tsuolbmajavri data in the light of our new Toskaljavri
results, following the replication rationale applied to the GISP2 and GRIP ice cores in Greenland (Dansgaard et al., 1993). A discussion about pollen stratigraphy and vegetation history of Toskaljavri will be presented elsewhere.

SITE DESCRIPTION

Toskaljavri (69°12'N, 21°28'E, 100.35 ha in size) is an arctic-alpine (oroarctic) lake situated in the most mountainous part of Finland (Fig. 1). It lies at 704 m a.s.l. and several surrounding peaks exceed 1000 m. The lake is located in a wide valley, and the shores are dominated by smooth relief, except in the west and south west where a steep slope reaches the shore. Bedrock consists mainly of so-called fjell schist, a strongly metamorphic rock. On the northern and northwestern side of the lake bedrock is dolomite. The lake is underlain partly by dolomite and partly by crystalline rock (National Board of Survey and Geological Society of Finland, 1990).

Climate is typically arctic alpine with a very short summer. Snow cover melts in June and ice cover disappears in late June or early July. The closest meteorological station is ca. 40 km southwest from the lake, but according to an interpolation-based Fennoscandian climate estimation model which takes into account the location and altitude of the site and distance from the sea, July mean temperature is ca. 9.7°C and January mean temperature is ca. −13.9°C. The interpolated value for annual precipitation is 415 mm (A. Odland, personal communication).

The surrounding vegetation consists mainly of low-alpine heaths. Common plants are Betula nana, Vaccinium myrtillus, Vaccinium vitis-idaea, Empetrum nigrum, grasses (Gramineae), and sedges (Carex spp.). However, the boundaries of the slightly and markedly oceanic vegetation sectors of northern Fennoscandia (Moen, 1999) are only 70 km and 130 km to the west, respectively, emphasizing the ecotonal position of Toskaljavri. The closest woodland consists of mountain birch (Betula pubescens ssp. tortuosa) located 2 km south of the lake. The limit of pine (Pinus sylvestris) forest is ca. 80 km to the
south and southeast but there are isolated pine forests on the Norwegian coast ca. 40 km to the west.

METHODS

Toskaljavri was sampled in April 1999 with a rodless piston corer (Chambers and Cameron, 2001). A 161-cm core was taken from the lake center where water depth is 21.5 m.

Pollen Data

The core was subsampled and analyzed at 1-cm intervals, giving a temporal resolution of ca. 60 yr. Pollen samples were prepared with standard KOH, HF, and acetolysis methods (Moore et al., 1991) and analyzed to the lowest possible taxonomic level. A minimum of 500 terrestrial pollen and spores were analyzed from each sample. Percentages of terrestrial pollen and spore taxa were calculated on the basis of their total sum. Percentages of aquatics were calculated on the basis of the total sum of terrestrial taxa plus aquatic taxa, and the percentages of Sphagnum on the basis of the total sum of terrestrial taxa plus Sphagnum. The pollen diagram was drawn with TILIA and TILIA.GRAPH (Grimm, 1990). Pollen and spore nomenclature follows Moore et al. (1991) with some exceptions: unidentified monolete spores are termed Dryopteris-type and oblong Cyperaceae pollen grains with large lacunae are called Carex-type. Plant taxonomy follows Hämet-Ahti et al. (1984).

Dating

Age control was established from eight accelerator mass spectrometry (AMS) 14C dates from plant macrofossils (Table 1). The dates were calibrated using Method A within CALIB 4.3 (Stuiver and Reimer, 1993) and the INTCAL 98 calibration data (Stuiver et al., 1998). An age–depth model (Fig. 2) for the core was developed using a nonparametric weighted regression model within the framework of generalized additive modeling (E. Heegaard and H. J. B. Birks, unpublished 2001). Modeling was carried out using seven calibrated dates as the basal date (159 cm) is considered too old on the basis of pollen stratigraphical correlations. The uppermost sediment was assumed to be modern and an age of \(-48 \pm 5\) yr B.P. was assigned to it for modeling. Ages for samples below 143 cm were estimated by extrapolation of the fitted model. Chronology is presented as calibrated years before present (cal yr B.P.).

Climate Reconstructions

For reconstructing July mean temperature (T\(_{\text{jul}}\)) and annual precipitation (P\(_{\text{ann}}\)), we used a modern pollen-climate data set of 113 samples from throughout Finland, 164 samples from throughout Norway, and 27 samples from northern Sweden. All 304 surface sediments (0–1 cm) are from small- to medium-sized lakes. Modern July mean temperatures and annual precipitation values were estimated for each lake by a standard interpolation and modeling procedure (see Seppä and Birks (2001) for details). Modern pollen-climate transfer functions based on all 156 pollen and spore taxa included in the calculation sum were developed using weighted-averaging partial least squares regression (ter Braak and Juggins, 1993). The resulting models have a good predictive ability, as assessed by leave-one-out cross validation (ter Braak and Juggins, 1993), with a root-mean-square error of prediction (RMSEP) of 0.99°C, an \(r^2\) between predicted and observed values of 0.71, and a maximum bias of 3.94°C for July mean temperature and a RMSEP of 341 mm, \(r^2\) of 0.71, and maximum bias of 993 mm for annual precipitation. See Seppä and Birks (2001) for further details.

RESULTS

A simplified pollen diagram is shown in Fig. 3 and reconstructed T\(_{\text{jul}}\) and P\(_{\text{ann}}\) values are given in Fig. 4. Both pollen data and loss-on-ignition analysis (LOI, not shown) indicate that the 87–79 cm section (ca. 5150–4700 cal yr B.P.) is anomalous. LOI suddenly drops from regular values of ca. 15% to ca. 5%, while the median mineral particle size is larger than above and below (R. Thompson, personal communication). The microfossil composition (Fig. 3) is characterized by decreasing levels of Pinus sylvestris, Juniperus communis, and Carex-type and by increasing levels of Lycopodium clavatum, Gymnocarpium dryopteris, and...
FIG. 3. Pollen diagram for Toskaljavri showing the most common pollen and spore taxa. The periods on the right-hand side refer to the three main climate periods as discussed in the text.

and Cryptogramma crispa. These features all suggest that the 87–79 cm sediment is secondary, most probably washed from the topsoil of the surrounding western steep slope by an avalanche or slush flow. This section was excluded from the climate reconstruction, which is based on the remaining 153 samples.

Early Holocene, 9600–8300 cal yr B.P. (Period I)

The basal assemblages are dominated by Betula, reaching maximum values of 60% in the lowermost sample. Ferns and lycopsids have values up to 10%. Pinus percentages increase.

FIG. 4. July mean temperature (A) and annual precipitation (B) reconstructions from Toskaljavri. The solid line is a LOESS smoother (span 0.25) fitted to the data to highlight the main trends. The section 5150–4700 cal yr B.P. (87–79 cm) is deleted from the reconstruction owing to assumed sediment redeposition.
gradually but there is a short period at 8700–8500 cal yr B.P. when values decrease. Before and during this decrease Juniperus has a peak of up to 20%. $T_{\text{jul}}$ values are below 12.0°C but rise steadily, in phase with increasing Pinus percentages. $P_{\text{ann}}$ is high, with values of ca. 600–800 mm, with some peaks reaching ca. 1000 mm.

**Early to Mid Holocene, 8300–6500 cal yr B.P. (Period II)**

After the drop at 8700–8500 cal yr B.P., Pinus values increase to ca. 60% Betula falls to below 30%. Ferns and lycopsinds decrease. Ericaceae-type and Vaccinium-type begin to occur with low but continuous values. This period represents the Holocene Thermal Maximum (HTM) or “climatic optimum.” Reconstructed $T_{\text{jul}}$ is above 12.0°C during the beginning of the period and reaches maximum values, ca. 12.4°C with some peaks up to 13.0°C, at 7500–7200 cal yr B.P. After this, the values start to decrease. $P_{\text{ann}}$ decreases slightly but stays above 500 mm.

**Mid to Late Holocene, 6500 cal yr B.P. to Present (Period III)**

Pinus has very high percentages at 6000–4000 cal yr B.P., with average values of 70–80%. Betula drops below 20% while Picea begins to occur at ca. 6000 cal yr B.P. Toward the top of the sequence Picea, Juniperus, Carex-type, and Gramineae increase while Pinus decreases. $T_{\text{jul}}$ declines to below 12.0°C at 6300 cal yr B.P. There is a period of slightly higher values at 4500–4000 cal yr B.P. Since then the cooling trend continues with some fluctuations, with the clearest rise occurring at 1300–1000 cal yr B.P. There is one very warm, probably anomalous, sample at 2500 cal yr B.P., where $T_{\text{jul}}$ exceeds 13.0°C. The last 800 cal yr B.P. are the coldest of the whole sequence, as reconstructed $T_{\text{jul}}$ falls to ca. 11.0°C at the top. $P_{\text{ann}}$ initially rises slightly but then decreases to a Holocene minimum at 3000–2000 cal yr B.P.

**DISCUSSION**

**Early-Holocene Climate**

In Fig. 5 the $T_{\text{jul}}$ record of Toskaljavri is compared with our $T_{\text{jul}}$ reconstruction from Tsuolbmajavri (Seppä and Birks, 2001), located in the same area but at lower altitude and having therefore higher reconstructed $T_{\text{jul}}$ values (Fig. 1). The following discussion is mainly based on these two reconstructions. For comparison, the $\delta^{18}$O record of the NGRIP Greenland ice core (Hammer, 2002; Johnsen et al., 2001) is also shown. Our results indicate that the early-Holocene climate in northern Fennoscandia was characterized by relatively low but steadily rising $T_{\text{jul}}$. In the Tsuolbmajavri record, which extends to ca. 9900 cal yr B.P., the values are below 11.0°C and rise to above 12.0°C at 9300 cal yr B.P. At Toskaljavri the record begins at 9600 cal yr B.P. with values of ca. 11.5–11.9°C from 9600 to 9000 cal yr B.P.

A common feature of many early-Holocene climate records from the North Atlantic region is a sudden cold period at ca. 8400–8000 cal yr B.P. which may have been caused by a transient perturbation in the North Atlantic oceanic circulation system (Alley et al., 1997; Klitgaard-Kristensen et al., 1998). Our records also indicate a slight cooling or at least a temporary slowing in the rising trend of summer temperatures in northern Fennoscandia at 8700–8000 cal yr B.P. The cooling is reflected palynologically by a transient decline of Pinus and a rise of Juniperus up to 15–20% (Fig. 3). A fall in $T_{\text{jul}}$ is evident only at Tsuolbmajavri (Seppä and Birks, 2001). The timing of this change agrees roughly with Barber et al. (1999), who propose that the event was triggered by catastrophic drainage of the Laurentide ice lakes in Canada at ca. 8470 cal yr B.P. and that the event lasted to ca. 8000 cal yr B.P. The relatively minor cooling in our records suggests that the event may have been an oceanic, North Atlantic event with a subdued impact in more continental areas of northern Europe. It is also possible, however, that the weak signal reflects the relative insensitivity of our pollen-stratigraphical record or that perturbations of the North Atlantic oceanic circulation did not significantly change summer temperatures but had a greater influence on other climatic variables in northern Fennoscandia (Nesje and Dahl, 2001).

The relatively low reconstructed summer temperature and high annual precipitation suggest a markedly oceanic climate during the early Holocene, until ca. 8500 cal yr B.P. (Seppä and Birks, 2001). The reconstructed early-Holocene climate with low summer temperatures, low seasonality, and high precipitation is supported by lake-level reconstructions from the Finnish tree-line area and northeastern Finland, which suggest that lake levels were high during the early Holocene and then experienced a major dry interval from ca. 8000 to 5800 cal yr B.P., after which levels again rose toward present levels (Hyvärinen and Alhonen, 1994; Sarmaja-Korjonen and Hyvärinen, 1999). In a larger, North Atlantic context the reconstructed climatic pattern for the early Holocene suggests that the climate until ca. 8500 cal yr B.P. was characterized by a stronger-than-present zonal circulation (Seppä and Birks, 2001), which resulted in increased cyclonic activity and enhanced penetration of moist Atlantic air flow across the Scandes mountains. Independent evidence for such a circulation is provided by an early-Holocene enrichment of $\delta^{18}$O in comparison to mid-Holocene values in carbonate sediments in Abisko, northern Sweden, suggesting a more vigorous flow of Atlantic air masses (Hammarlund et al., 2002).

The interpretation of a cool, moist climate contrasts with early-Holocene solar radiation patterns (Fig. 6). Due to the summer perihelion and greater tilt of the Earth’s axis, early-Holocene radiation conditions were more extreme than at present and the amount of solar radiation received in July at the atmosphere surface at ca. 10,000 cal yr B.P. at 70°N was ca. 47 W/m² higher than at present (Berger, 1978). We suggest that the low $T_{\text{jul}}$ of Tsuolbmajavri record at 9900–9500 cal yr B.P. reflects at least partly the decreasing periglacial influence of the Scandinavian ice sheet and that until ca. 8200 cal yr B.P. the effect of high summer solar radiation was subdued by the cooling effect of enhanced Atlantic air flow.
The atmospheric pressure patterns of northern Fennoscandia have strong teleconnections to the climate over the Barents Sea. According to the model of Ádallsvik and Loeng (1991), predominantly low pressure conditions over northern Fennoscandia would result in an enhanced penetration of the North Atlantic Current to the Barents Sea. The occurrence of fossils of two thermophilous molluscs, Zirphaea crispata and Modiolus modiolus, on Svalbard at 9700–8800 cal yr B.P. accords with our records, suggesting a maximal supply of warm Atlantic surface water to Svalbard during the early Holocene (Salvigsen et al., 1992). The occurrence of the richest and most abundant plant macrofossil assemblages during the early Holocene (ca. 9000 cal yr B.P.) on Svalbard (Birks, 1991) can similarly be linked to the climatic effect of a more intense flow of the Atlantic waters.

Holocene Thermal Maximum

After the minor cooling at 8600–8200 cal yr B.P., $T_{\text{jul}}$ rose to ca. 13.0$^\circ$C at Tsuolbmajavri while there was a steady increase to 12.1$^\circ$–12.3$^\circ$C at Toskaljavri. These constitute the Holocene Thermal Maximum in our records. Comparisons with the reconstructed modern $T_{\text{jul}}$ values indicate that during the HTM, summer temperatures in northern Fennoscandia were roughly 1.6$^\circ$–1.8$^\circ$C higher than today. The timing of the HTM to 8000–6500 cal yr B.P. in the Toskaljavri and Tsuolbmajavri reconstructions (Fig. 5) is roughly synchronous with the NGRIP $\delta^{18}$O record, which suggests maximum temperatures at 8000–7200 cal yr B.P.

The occurrence of maximum summer temperatures at 8000–6500 cal yr B.P. was probably due to at least three complementary factors. First, the Fennoscandian ice sheet melted ca. 10,000–9500 cal yr B.P. (Lundqvist, 1991) and it no longer affected the climate of northern Fennoscandia. Second, although the amount of summer solar radiation at 70$^\circ$N was lower than during the early Holocene, it was still ca. 40 W/m$^2$ higher than at present (Berger, 1978). Third, and related to this, there may have been a shift from the dominantly zonal atmospheric circulation to more stable high-pressure summer conditions over northwestern Eurasia with the development of persistent blocking anticyclones and a more southerly and southeasterly air flow (Seppä and Birks, 2001).

At present, the longest and warmest summers in northern Fennoscandia occur in connection with such deep and warm anticyclones, which often bring calm weather and cloudless skies (Rex, 1950; Johannessen, 1970). During such periods temperature conditions are extremely influenced by the radiation balance.
of the Earth’s surface (Johannessen, 1970). The blocking anticyclonic pattern is particularly unfavorable to glacier growth both in summer and winter, in contrast to strong zonal circulation, which enhances precipitation, snow accumulation, and glacier growth on the Scandes (Rex, 1950). The decrease or disappearance of many glaciers in Fennoscandia between 8000 and 6000 cal yr B.P. provides independent support for the proposed atmospheric pattern (Nesje and Dahl, 1993; Dahl and Nesje, 1994; Nesje et al., 1994; Karlén, 1998). The blocking action would also weaken Atlantic inflow to the Barents Sea (Ádalansvik and Loeng, 1991) and would explain the lower reconstructed sea-surface temperatures for the northern Norwegian Sea (Hald and Aspeli, 1997) and the absence of thermophilous molluscs on Svalbard ca. 8000–6000 cal yr B.P. (Salvigsen et al., 1992).

Mid- to Late-Holocene Cooling

Both pollen-based records suggest gradual cooling of $T_{\text{jul}}$ since ca. 6500 cal yr B.P. (Fig. 5). They also indicate fluctuations in $T_{\text{jul}}$, but most of the fluctuations have occurred asynchronously between sites, which, assuming that the age models are reliable, suggests that the changes may not reflect real climatic events. The clearest consistency between records after the HTM is during the last 2000 cal yr B.P. where our reconstructions agree with the traditional concept of a “Medieval Warm Period” (MWP) and “Little Ice Age” in the North Atlantic region (Dansgaard et al., 1975) and in northern Fennoscandia (Korhola et al., 2000). According to Lamb (1997), the MWP in parts of the Arctic began at ca. AD 300–400 and came to an end ca. AD 1300, i.e., roughly synchronously with our data. Lamb (1997) estimates that during the warmest phase of the MWP at AD 1000–1200 summer temperatures in England were 0.7–1.0°C warmer than during the 20th century, which is very close to our reconstructions. There is also a clear correlation between our MWP reconstruction and several records from Greenland ice cores. Comparisons (Fig. 7) of a smoothed $T_{\text{jul}}$ record from Toskaljavri with measured borehole temperatures of the GRIP and Dye 3 ice cores (Dahl-Jensen et al., 1998) and the $\delta^{18}$O record from the Crête ice core (Dansgaard et al., 1975) show the strong similarity in timing of the MWP between the records.

CONCLUSIONS

The main climatic trends of our two pollen-based climate reconstructions suggest a gradual early-Holocene rise of summer temperatures and a gradual mid- and late-Holocene cooling with no unambiguous indications for abrupt climatic shifts. This accords with the frequently presented view that the Holocene has been characterized by a relatively stable climate with fluctuations of a lower amplitude than those during the Weichselian (Dansgaard et al., 1993; McManus et al., 1994; Fronval and
Jansen, 1997). The model of gradual temperature change is not, however, concordant with a number of recent contributions from the North Atlantic area suggesting short-term cold shifts that may have occurred in cycles of ca. 1500 yr (Bond et al., 1997). Our data suggest that if such cold periods have occurred they have not been of the magnitude to be detectable with our relatively coarse reconstruction method and are thus masked in the long-term vegetational and climatic change. Alternatively they may reflect local or regional climatic or oceanographic changes with little or no continental-scale impacts.

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