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Abstract: July mean temperature and annual precipipation during the last 9900 cal. yr BP were reconstructed from pollen assemblages preserved in a sediment core from northern Finland. Quantitative reconstructions were performed using a modern pollen-climate calibration model based on weighted-averaging partial least squares regression. The predictive ability of the model was evaluated against modern meteorological data using leave-one-out cross-validation. The prediction error for July mean temperature is *c*. 1.0°C and for annual precipitation 340 mm. The July mean temperatures during the earliest Holocene were low, *c*. 11.0°C, and annual precipitation was high, *c*. 600–800 mm. Between 8200 and 6700 cal. yr BP July mean temperatures reached their maxima, 12.5–13.0°C, which are *c*. 1.4–1.7°C higher than at present. At the same time precipitation decreased. During the late Holocene, July mean temperatures declined and the last 2000 years have been the coolest since the early Holocene. Precipitation has slightly increased. The spatial coherence between our results and of several other climate reconstructions from northern Europe indicates that the Holocene climate was strongly influenced by North Atlantic oceanic and atmospheric circulation patterns. We propose that the distinctly oceanic climate of the early Holocene was due to enhanced westerly (latitudinal) airflow which was replaced at *c*. 8200 cal. yr BP by a more meridional flow pattern and by the development of predominantly anticyclonic summer conditions.

Key words: Pollen, climate, quantitative reconstructions, Fennoscandia, North Atlantic, atmospheric circulation, Holocene.

Introduction

Continuous Holocene climate records from the North Atlantic area have been obtained from deep-sea cores and from Greenland ice cores. On a millennial timescale the physical and chemical data from the GRIP and GISP2 ice cores from Greenland suggest that the climate over Greenland and the North Atlantic during the Holocene Climatic Optimum or Hypsithermal *c*. 8000–5000 cal. yr BP was 2–3°C warmer than at present (Stuiver *et al*., 1995; Dahl-Jensen *et al*., 1998) and that the cooling towards modern conditions since *c*. 5000 cal. yr BP occurred gradually (Dahl-Jensen *et al*., 1998). The deep-sea sediment evidence in general indicates that sea-surface temperatures (SST) have been relatively stable during much of the Holocene and that the Nordic Sea has been characterized by relatively stable oceanographic conditions (Fronval and Janssen, 1997). However, there are several inconsistencies between climate proxies based on various records from the ice cores and between the ice core data and deep-sea core data. The glaciochemical time-series from the GISP2 core and some deep-sea core records, for example, suggest that the generally stable Holocene climate has been punctuated by frequent cold periods (O'Brien *et al*., 1995; Bond *et al*., 1997; Bianci and McCave, 1999).

From the north European viewpoint, the climatic history of the North Atlantic is of great importance because the North Atlantic has a direct influence on the climate of northwest Europe. This

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influence mainly results from the upper troposphere jet-stream, entering the Eurasian continent between 65°N and 50°N, and the associated strong westerly atmospheric circulation (Wallén, 1970). Superimposed on the westerly air flow is the heat and moisture transport from the North Atlantic Current (the Gulf Stream). Due to these strong teleconnections between ocean and continent, long-term air-temperature fluctuations in Fennoscandia are closely connected with atmospheric circulation variations (Johannessen, 1970). Consequently, palaeoclimatic reconstructions from northwest Europe may serve not only to elucidate past regional climate variations but also to provide insights into Holocene atmospheric and oceanic circulation dynamics in the North Atlantic/Nordic Sea area.

The understanding of the dynamics and interactions of the changing insolation and atmospheric circulation patterns during the Holocene requires precise continuous climate records from sensitive continental sites in northern Fennoscandia. Such data can be best provided by ultra-high-resolution studies which have been carried out in northern Fennoscandia using dendrochronological data (e.g., Briffa *et al*., 1992; Zetterberg *et al*., 1996). However, the dendrochronological data are most suited for decadal to centennial scale climate reconstructions and to detect high-frequency climate events. Tree-ring-based reconstructions over longer timescales may be less reliable due to the problems resulting from standardization techniques (Cook *et al.*, 1995; Briffa *et al.*, 1996; Bradley, 1999) and possible genotypic plasticity of the tree species in response to low-frequency climate changes.

One potential source for low-frequency, centennial to millennial scale climate reconstructions in northern Fennoscandia is provided by pollen-stratigraphical data. Since the pioneering work by Webb and Bryson (1972), pollen stratigraphies in North America and central Europe have been widely used for quantitative climate reconstructions. Several numerical techniques have been used, mainly based on pollen response surface, inverse regression, or modern analogue techniques (e.g., Guiot, 1990; Bartlein and Whitlock, 1993; Birks, 1995). Statistical developments in quantitative palaeolimnology (e.g., ter Braak and Juggins, 1993; ter Braak *et al*., 1993; ter Braak, 1995) have shown that techniques based on weighted averaging regression and calibration, particularly weighted-averaging partial least squares (WA-PLS) regression, have several advantages in environmental reconstruction. Besides high predictive ability and low bias, as assessed by statistical cross-validation (Birks, 1995), the major advantage is the ability of WA-PLS to extrapolate, to some degree, possible 'no-analogue' situations as may occur in, for example, the early Holocene (ter Braak, 1995; Birks, 1998).

Here we present quantitative pollen-based July mean temperature (T_{jul}) and annual precipitation (P_{ann}) reconstructions based on WA-PLS from Tsuolbmajavri in northern Finland, a small lake that is located at the ecotone between the Boreal forest and Arctic-alpine ecosystems and at the boundary between the oceanic and continental climate regimes. The site is assumed to be sensitive in climatological terms, especially to shifts in the dominance of atmospheric circulation types. For quantitative climatic reconstructions we use a modern Fennoscandian pollen-climate calibration data set, where particular attention has been paid to consistency in site selection, methodology and pollen taxonomy, and to model estimation and selection using statistical cross-validation.

Study area

Tsuolbmajavri is located at 68°41′30″N, 22°05′E, at an elevation of 526 m above sea level (a.s.l.) (Figure 1). Lake size is 14 ha and the maximum depth is 5.35 m. It has a small inlet in the north and a small outlet in the south. A rolling bedrock-dominated

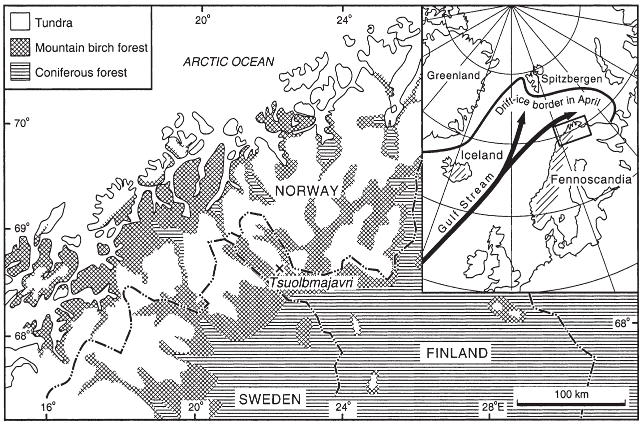

Figure 1 The location of Tsuolbmajavri in relation to the major vegetational regions in northwestern Fennoscandia. The inset map shows the current role of the Gulf Stream and sea ice in influencing the regional climatology of the area.

landscape with the highest peaks over 650 m a.s.l. surrounds the lake. The dominant bedrock types are quartzite, quartz diorite, and granodiorite. The Tsuolbmajavri area, as the whole of northern Fennoscandia, was ice-covered during the Weichselian glaciation. Deglaciation in the area took place from north to south (Lundqvist, 1991) and Tsuolbmajavri is thought to have been deglaciated *c*. 9400 ^{14}C yr BP (*c*. 10 300 cal. yr BP) (Lundqvist, 1991).

Although the closest meteorological station to Tsuolbmajavri is located at an altitude of 320 m a.s.l. and 33 km to the south of the lake, the present climate in the area can be estimated from 1961–90 Climate Normals data from nearby meteorological stations (two in Norway, five in Finland, four in Sweden), using a linear interpolation model which takes into account the altitude of site and its distance from the sea (A. Odland, personal communication; Olander *et al*., 1999). According to this model, the present T_{jul} at the lake is estimated to be 11.0°C, the January mean temperature –14.5°C, and P_{ann} 450 mm yr^{-1}. These values are typical for the area located on the leeside of the Scandes Mountains and, consequently, for the area where westerly air flow has only limited climatic influence today.

Biogeographically, Tsuoblmajavri lies at the ecotone between Boreal forest and the (oro)Arctic tundra (Figure 1). It is surrounded by forest dominated by mountain birch (*Betula pubescens* ssp. *tortuosa*), the northernmost variant of the northern Boreal zone in Fennoscandia (Hämet-Ahti, 1963), but the upper tree-line of mountain birch is only *c*. 20 m above the altitude of the lake. Above that, treeless low-alpine heaths prevail. The northern limit of continuous pine (*Pinus sylvestris*) forest is *c*. 30 km to the south and southeast of the lake, whereas the northern limit of spruce (*Picea abies*), the other conifer tree species in northern Fennoscandia, is *c*. 80 km to the south and southeast.

In addition to mountain birch, the flora of the birch forest is characterized by shrubs and dwarf shrubs. *Juniperus communis*, *Vaccinium myrtillus* and *V. vitis-idaea* are typical field-layer species. A patch of lush tall-herb vegetation occurs at the southern end of the lake. Peatland covers the northern margin of the lake. There the dominant species are *Salix lapponum*, *Carex* spp., *Rubus chamaemorus*, *Potentilla palustris*, *Equisetum* spp. and *Sphagnum* spp. Tall juniper bushes characterize the heath vegetation directly above the mountain birch tree-line. Above that, the vegetation consists of dry, oligotrophic heaths, where *Empetrum nigrum*, *Betula nana*, *Vaccinium uliginosum* and *V. myrtillus* are frequent. Soils in the valley bottom are thick and humic but above the tree-line thin inceptisols prevail. Further details of the site and its surroundings are given by Seppä and Weckström (1999).

Material and methods

Pollen data

The pollen samples were prepared from a 291 cm long sediment core, taken from the centre of Tsuolbmajavri. Samples were prepared with standard KOH, HF and acetolysis techniques at a 2 cm sample interval. *Lycopodium* tablets were added to allow pollen concentration and influx estimations (Stockmarr, 1971). Pollen identifications follow the harmonized identification procedures and taxonomy of the NORD-CHILL project (S.M. Peglar, unpublished), mainly based on Moore *et al*. (1991) and modern pollen and spore reference collections. The 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 the TILIA and TILIA.GRAPH programs (Grimm, 1990). Pollen nomenclature follows Moore *et al*. (1991) with some exceptions: Polypodiaceae spores of Moore *et al*. (1991) are termed *Dryopteris*-type and oblong Cyperaceae pollen grains with large luminae are called *Carex*-type. Plant taxonomy follows Hämet-Ahti *et al*. (1984).

Chronology

Fourteen AMS radiocarbon dates provide the chronological control for the core (Table II in Seppä and Weckström, 1999). Four of the dates are based on macroscopic remains of *Drepanocladus* spp. moss and 10 on bulk clay gyttja. The sediment contained no terrestrial plant macrofossils which could be used for dating.

Age-depth models were developed using calibrated radiocarbon dates. The radiocarbon dates were calibrated using the bidecadal tree-ring data set A and method A in the CALIB 3.03c program (Stuiver and Reimer, 1993) and a 10-sample curve smoothing (Törnqvist and Bierkens, 1994). As the basal date based on clay-gyttja is of lateglacial age (10940 ± 95 ^{14}C yr BP) and is clearly too old (Seppä and Weckström, 1999) in relation to the deglaciation history of the area (Lundqvist, 1991), it was excluded from the age-depth modelling. The remaining dates indicate roughly linear sedimentation rate and the four dates from moss are consistent with the other nine dates and do not deviate from this general linear pattern.

Age-depth models based on the 13 dates were developed by non-parametric weighted regression within the framework of generalized additive models (Heegard and Birks, unpublished data). Three different variance functions (constant variance, variance proportional to the mean, variance proportional to (mean)^2), an identity link function, a cubic spline incorporating different degrees of freedom or ‘roughness’ were used, and the uppermost sediment forced to an age of –46 cal. yr BP. The simplest parsimonous age-depth model was then selected on the basis of the model being statistically significant, having the fewest terms in the spline smoother, and showing no serious deviations or patterns in the resulting regression diagnostic plots. The simplest statistically significant model (Figure 2) is based on a variance function proportional to the mean and with a three degree-of-freedom equivalent smoother parameter. This basically linear model is adopted here. Approximate ages for the basal 20 cm were estimated by linear extrapolation of the fitted regression model. All ages (cal. yr BP) quoted for Tsuolbmajavri in the text are based on this age-depth model. In the discussion, where comparisons are made with radiocarbon dated events at other sites in northern Europe, the ^{14}C ages have been calibrated directly using CALIB 3.03, method A, and the statistical tree-ring data set A.

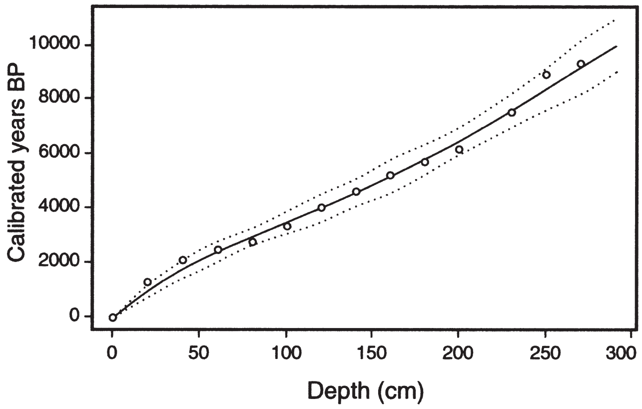

Figure 2 The fitted age-depth model for Tsuolbmajavri based on 13 calibrated radiocarbon dates and the assumption that the uppermost sediment is contemporary. The solid line is the model fitted by a spline smoother and the dotted lines are the 95% confidence intervals for the fitted model.

Modern pollen-climate data set

The modern pollen data set consists of 113 modern surface-sediment samples from Finland, collected in the summer of 1997, 164 samples from Norway, collected between 1991 and 1994, and 27 samples from northern Sweden, collected in 1993–94. The location of the sites is shown in Figure 3. All the samples are from small to medium-sized lakes and were taken with a gravity corer from a rubber boat from the deepest part or centre of the lakes. The top 1 cm of sediment was taken for the surface sample. In the selection of lakes, lakes were avoided that were in exceptional soil and/or vegetation areas (e.g., fluvioglacial formations in Finland) or extensively cultivated areas, or had large marginal sedge swamps.

The pollen samples were analysed to the lowest possible taxonomic level. The Finnish samples were analysed by H.S. and the Norwegian and Swedish samples by S.M. Peglar. The pollen-

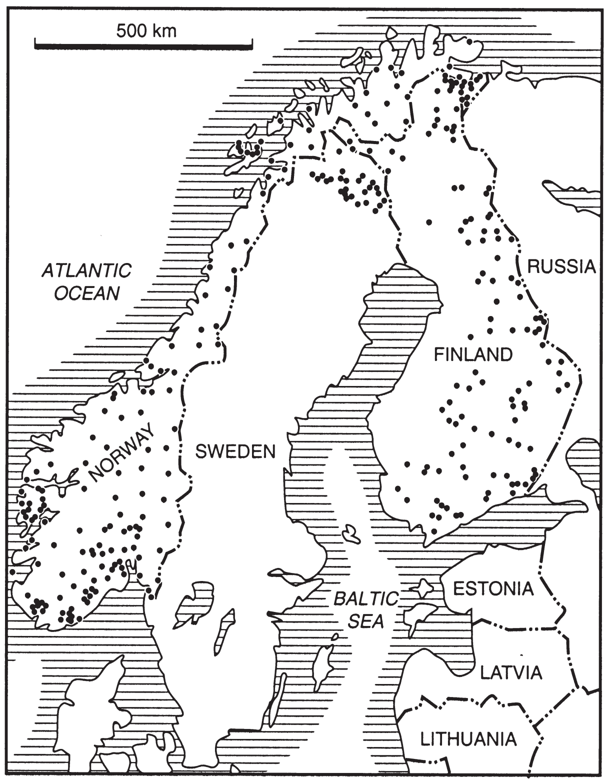

Figure 3 The position of the 304 modern pollen samples from Norway (164), Finland (113) and northern Sweden (27), used in the modern pollen-climate calibration function.

taxonomic compatibility between the two analysts was improved by three taxonomic sessions during which problematic taxa were studied and by cross-counting the same pollen samples and comparing the results statistically. The results of this analytical quality control will be published elsewhere. In brief, they indicate that analyst-dependent identification and counting differences are a statistically significant factor, but that the difference between the two analysts involved in this study is small and barely significant statistically.

Modern July mean temperatures (T_{jul}) and annual precipitation values (P_{ann}) were estimated by Arvid Odland for each of the 304 samples using the 1961–90 Climate Normals data from grids of nearby meteorological stations in Norway, Sweden and Finland. P_{ann} values were estimated by interpolation between meteorological stations, with allowance being made for elevation. T_{jul} values were estimated by applying a lapse rate of 0.57°C per 100 m altitude (Laaksonen, 1976) to allow for temperature changes along elevational gradients and by applying empirically derived regional west–east lapse rates to allow for temperature changes with increasing distance from the coast for sites west of the Scandes. T_{jul} for each such lake was estimated from their distance from the coast and their elevation. For sites east of the Scandes, temperatures were estimated by interpolation from the nearest meteorological stations, with a lapse-rate correction for altitude.

Modern pollen-climate transfer functions were developed using weighted-averaging partial least squares (WA-PLS) regression (ter Braak and Juggins, 1993; ter Braak *et al.*, 1993; Birks, 1995; ter Braak, 1995). All 156 pollen and spore taxa included in the calculation sum were used in the transfer function. Their percentages were transformed to square roots in an attempt to optimize the 'signal-to-noise' ratio and to stabilize the variances. WA-PLS was selected because it has been shown in many empirical and several theoretical studies to perform as well as or even better than other regression and calibration procedures commonly used to develop organism-environmental transfer functions (see ter Braak *et al.*, 1993; ter Braak, 1995; Birks, 1995; 1998).

The performance of the WA-PLS transfer functions is reported here (Table 1) as the root mean square error of prediction (RMSEP), the coefficient of determination (r^2) between observed and predicted values, and the maximum bias (ter Braak and Juggins, 1993), all based on leave-one-out cross-validation or jack-knifing (ter Braak and Juggins, 1993; Birks, 1995). Two-component WA-PLS models were selected (Table 1) on the basis of low RMSEP, low maximum bias and the smallest number of 'useful' components (Birks, 1998). Plots of the predicted (in leave-one-out cross-validation) T_{jul} and P_{ann} and observed T_{jul} and P_{ann} and of the residuals (predicted-observed values) and observed values are given in Figure 4. They show that the pollen-climate transfer function overestimates T_{jul} below about 10°C (Figure 4C), presumably because of far-blown pollen from the lowlands being incorporated in the surface sediments of lakes above the tree-line where the July mean temperature is less than 10°C. There is no strong bias in the annual precipitation model (Figure 4B), except for the tendency to underestimate P_{ann} at sites that receive more than 2000 mm P_{ann}. Full details of the modern pollen-climate data sets will be published elsewhere.

In the plot of residuals (Figure 4) and in the stratigraphical plots of the inferred T_{jul} (Figures 6 and 7) and P_{ann} (Figure 7) a LOESS scatter plot smoother (span = 0.25; order = 1) (Cleveland, 1979) has been fitted to help highlight the major trends.

Table 1 Performance statistics for weighted-averaging partial least squares (WA-PLS) regression models for the modern pollen-climate data set (304 samples) from Finland, Norway and northern Sweden. Root mean square error of prediction (RMSEP), coefficient of determination (r^2), and maximum bias are given based on leave-one-out cross-validation for 1-, 2- and 3-component WA-PLS models for July mean temperature and annual precipitation. The 2-component models were selected for reconstruction purposes because of their low RMSEP, low maximum bias and small number of components

WA-PLS component	RMSEP	r^2	Maximum bias
July mean temperature (range = 7.7–17.1°C, mean = 14.5°C, median = 13.3°C)			
1	1.084°C	0.669	4.170°C
2	0.995°C	0.722	3.939°C
3	0.992°C	0.726	3.636°C
Annual precipitation (range = 300–3234 mm, mean = 912 mm, median = 651 mm)			
1	364.35 mm	0.675	1210.15 mm
2	341.31 mm	0.706	993.69 mm
3	361.49 mm	0.675	960.15 mm

Results

Only the general outlines of the pollen stratigraphy and inferred vegetation history are presented here as a full description and discussion of the pollen-stratigraphical results, including pollen-influx data, are presented in Seppä and Weckström (1999). A summary pollen diagram is shown in Figure 5.

Early Holocene *c.* 9900–8200 cal. yr BP

The pollen assemblage is dominated by *Betula* whose maximum values are above 60%. *Lycopodium clavatum*, *Diphasiastrum*, *Lycopodium annotinum*, *Lycopodium* undiff., *Gymnocarpium dryopteris* and *Dryopteris*-type spores occur consistently with significant values. These features suggest that the recently deglaciated terrain supported an open birch forest. The field-layer vegetation was relatively diverse with a significant component of lycopods and ferns. *Juniperus communis* pollen has a distinct peak at *c.* 8700 cal. yr BP. The values of *Pinus sylvestris* pollen increase throughout this period.

The reconstructed July temperatures imply low but rapidly rising T_{jul} (Figure 6). They are estimated at *c.* 11.0°C at the bottom of the core and stay below 12.5°C during the whole interval. The rising trend is clearly connected to the fall of *Betula* and to the increase of *Pinus sylvestris* pollen percentages. P_{ann} figures are highest for the whole Holocene in this period (Figure 6). The lowest five reconstructed values (before the rise of *Gymnocarpium dryopteris* and *Dryopteris*-type spores) are below 450 mm yr^{-1}, after which (*c.* 9550 cal. yr BP) the reconstructed P_{ann} rises rapidly to above 600 mm yr^{-1} with maximum values above 800 mm yr^{-1}. They stay above 600 mm yr^{-1} to 7000 cal. yr BP.

Mid-Holocene *c.* 8200–5700 cal. yr BP

Pollen assemblages are characterized by the continuous rise of *Pinus sylvestris* percentages. Maximum percentages are over 60% and suggest the local presence of a sparse pine forest in the surroundings of Tsuolbmajavri at least from *c.* 6900 to 5700 cal. yr BP. *Betula* and most of the other typical early-Holocene taxa decrease. Values of *Empetrum nigrum*, Ericaceae-type and *Vaccinium*-type rise at *c.* 7100 cal. yr BP, probably reflecting a change in the field-layer vegetation and the development of a more acidic soil following the expansion of pine.

The climate reconstructions suggest that Holocene T_{jul} maxima were achieved during this period. The reconstructed values rise steadily and generally exceed 13.0°C from *c.* 7950 to 6750 cal. yr BP. They then begin to fall but are still above 12.0°C at *c.* 5700 cal. yr BP. P_{ann} rapidly falls and is less than 500 mm yr^{-1} *c.* 6650

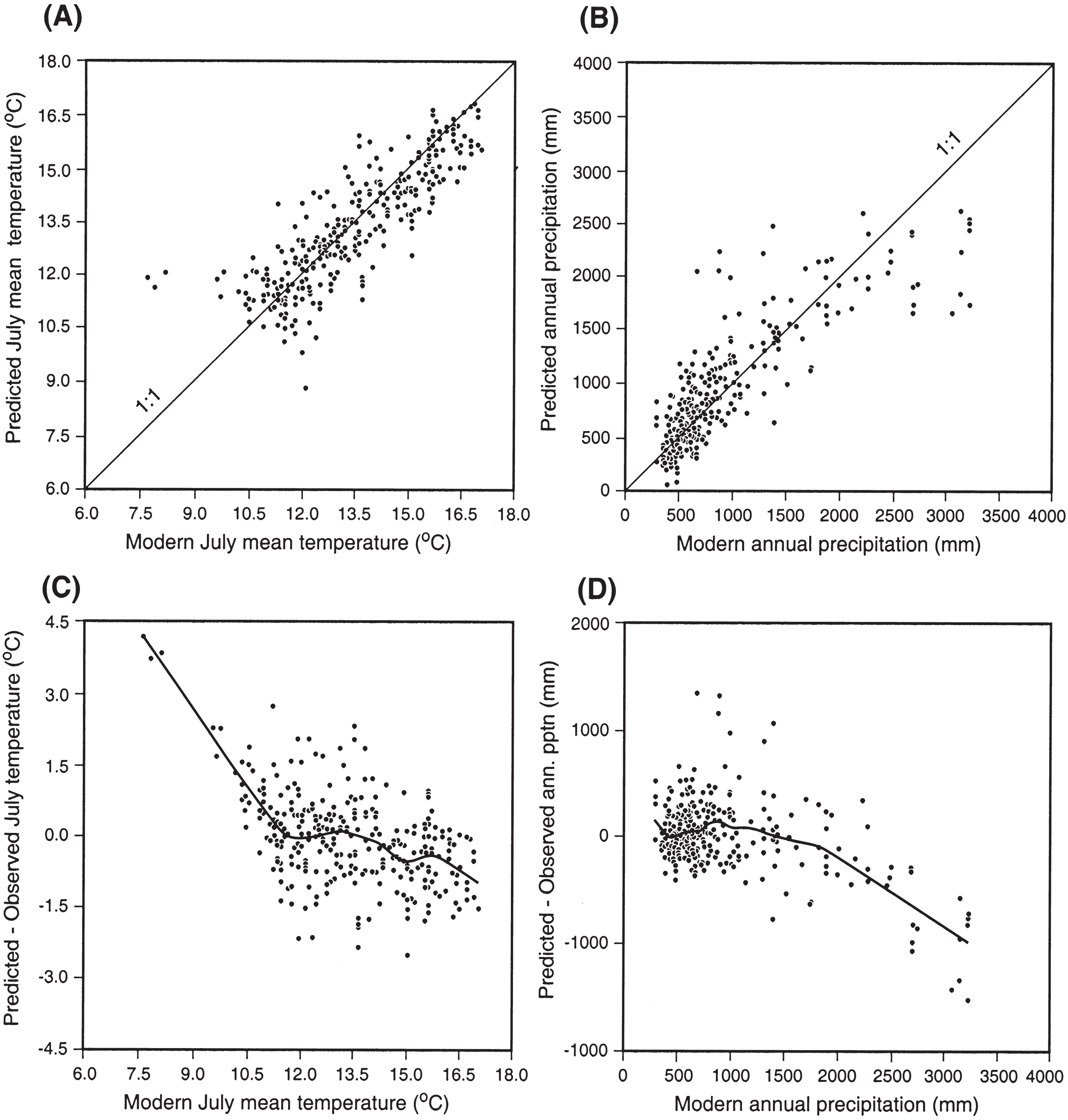

Figure 4 Scatter plots of: (A) predicted July mean temperature against modern July mean temperature; (B) predicted annual precipitation against modern annual precipitation; (C) residuals (predicted-observed) against modern July mean temperature; (D) residuals against modern annual precipitation. All the predicted values are based on leave-one-out cross-validation. In the upper plots (A and B), a 1:1 line is drawn; in the lower plots (C and D) the fitted line is a LOESS smoother (span = 0.25; order = 1).

to 5700 cal. yr BP. This decrease is mainly generated by the increasing *Pinus sylvestris* pollen percentages and by the decrease of lycopod and fern spore values.

Mid- to late Holocene *c.* 5700–2400 cal. yr BP

Pinus sylvestris is the dominant pollen taxon with steady values of *c.* 40–50%. *Picea* percentages start to rise at *c.* 3550 cal. yr BP and have their maximum values above 5% at *c.* 3150 cal. yr BP, while *Betula* pollen values decrease slightly.

Reconstructed T_{jul} is below 12.0°C at first. There is a period of remarkably steady values between *c.* 5050 and 3950 cal. yr BP during which T_{jul} stays at *c.* 12.0°C. After this, the values rise slightly and are above 12.5°C from *c.* 3400 to 3100 cal. yr BP but start to fall again. P_{ann} rises firstly to above 500 mm yr^{-1} but then decreases again and reaches the Holocene minimum values, below 400 mm yr^{-1}, between *c.* 4250 and 3400 cal. yr BP. After this, the values begin to increase towards the present.

Late Holocene 2400 cal. yr BP to present

A distinct decrease of *Pinus sylvestris* pollen percentages starts at *c.* 2400 cal. yr BP. This reflects the gradual retreat of the pine tree-line to its modern position. At the top of the core, *P. sylvestris* values are below 30%. Percentages of *Juniperus communis*, *Carex*-type, and Gramineae pollen increase suggesting some opening of the vegetation. *Sphagnum* values increase gradually during the late Holocene.

T_{jul} is clearly falling in step with the falling *Pinus sylvestris* percentages. It drops below 12.0°C at *c.* 1900 cal. yr BP, and then,

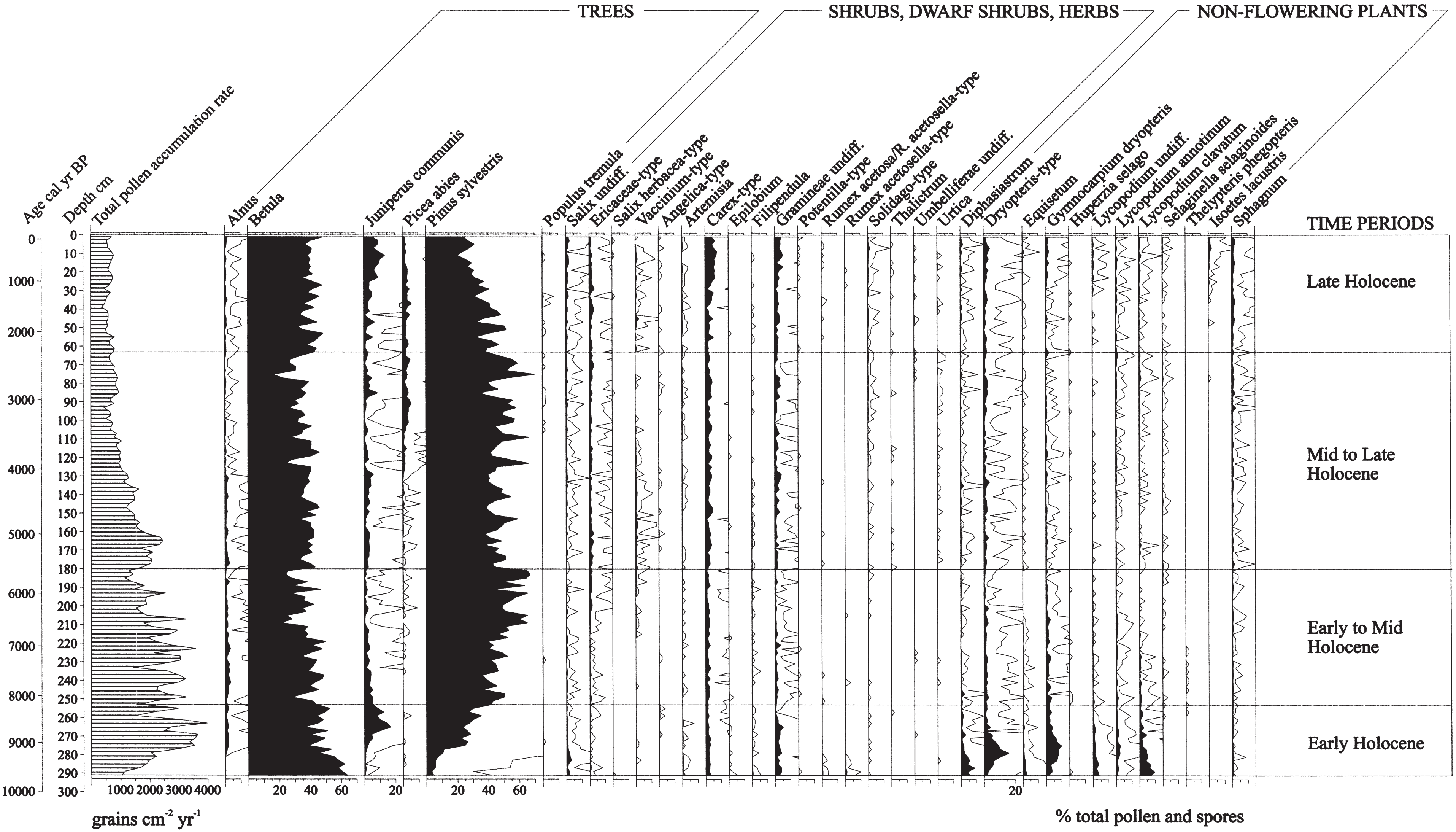

Figure 5 Summary pollen diagram from Tsuolbmajavri. The age scale in modelled calibrated years BP is shown along with the four phases discussed in the text. The total pollen- and spore-accumulation rate (grains·cm^{-2}·yr^{-1}) is also shown. The hollow silhoutte curves denote the 10 × exaggeration of the percentages.

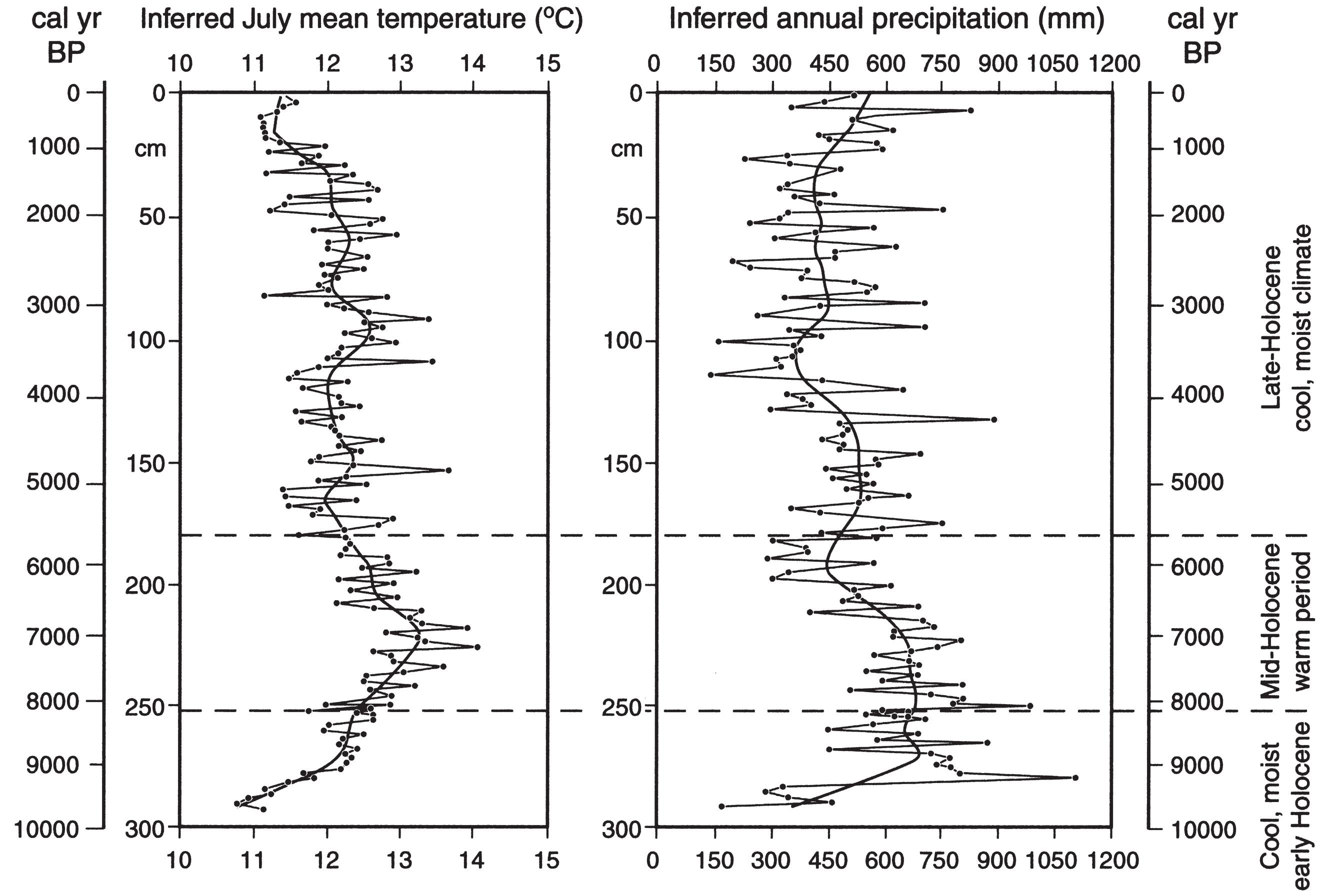

Figure 6 Inferred July mean temperature and annual precipitation at Tsuolbmajavri. The values are derived by applying a 2-component WA-PLS model (see Figure 4 and Table 1) to the full pollen-stratigraphical data at Tsuolbmajavri. The individual samples are joined and the fitted line is a LOESS smoother (span = 0.25; order = 1) to highlight the major long-term trends. The three major climatic phases discussed in the text are also shown.

for the first time since the early Holocene, decreases to below 11.5°C, at *c.* 800 cal.BP. The reconstructed present-day T_{jul} is 11.6°C which is 0.6°C higher than the estimated modern value. P_{ann} rises during the period, presumably due to the falling *Pinus sylvestris* values and slightly rising values of *Dryopteris*-type and *Gymnocarpium dryopteris* spores. The highest P_{ann} figures during the second millennium AD are above 700 mm yr^{-1}. The reconstructed modern P_{ann} is 590 mm yr^{-1} while the estimated modern value based on interpolated meteorological data is 450 mm yr^{-1}.

Discussion

There are several limitations and potential problems in the pollen-stratigraphical data, chronology and resulting climatic reconstructions from Tsuolbmajavri. A total of 145 samples were pollen-analysed at the site, giving a sample resolution of about 70 calibrated years. As commonly occurs with finer and finer pollen-stratigraphical resolution, there is considerable variation in the pollen percentages from sample to sample (Figure 5) even though at least 500 grains were counted in each sample. This variability inevitably results in considerable variation from sample to sample in the climate reconstructions (Figure 6). The sample-specific errors of reconstruction, as estimated by leave-one-out cross-validation and Monte Carlo simulation (Birks, 1995) are about ±1.0–1.1°C for July mean temperature and about ±350–370 mm for annual precipitation.

It is unclear what is 'signal' and what is 'noise' in the basic pollen-stratigraphical data and what is related to inherent counting errors, within-lake taphonomic and sedimentary processes, and 'real' changes in pollen composition. The LOESS smoothers (Figures 6 and 7) are used to identify and highlight the major, long-term trends in the climate reconstructions. They are not designed to pick up fine-scale changes. While it is tempting to match particular peaks or troughs in the climate reconstructions with short-term climatic changes known from elsewhere in the North Atlantic area, we have avoided this temptation because of our inability to distinguish between 'real' short-term changes based on one or two samples and inherent sample-to-sample variation and because of the insecure chronology at Tsuolbmajavri.

Dating control is not good at our site because it is based on AMS dating of bulk limnic sediment or *Drepanocladus* moss. The dates may thus be liable to some hard-water effects (Oldfield *et al.*, 1997; Barnekow *et al.*, 1998). Age-depth models can never be better than the radiocarbon dates and their calibrated equivalents on which they are based, and our age model may thus be in error. Even if we assume that the radiocarbon dates are all reliable, the resulting age-depth model has unavoidable uncertainties, arising from the counting errors in the radiocarbon assay, from the statistical problems in radiocarbon calibration procedure, and from the statistical errors within the model. The resulting uncertainties (one standard deviation) in the age-depth model (Figure 2) range from about 150 years at 1500 cal. yr BP to about 500 years at 9000 cal. year BP. These dating problems and the age-depth model uncertainties become particularly critical when we compare estimated ages from different proxy records based on radiocarbon dating of wood, peat and lake sediments and on ice-core chronologies.

In the light of these dating uncertainties, the magnitude of the sample-specific reconstruction errors, and the potential insensitivity of pollen percentages to fine-scale climate change, we restrict our interpretations to long-term centennial and millennial

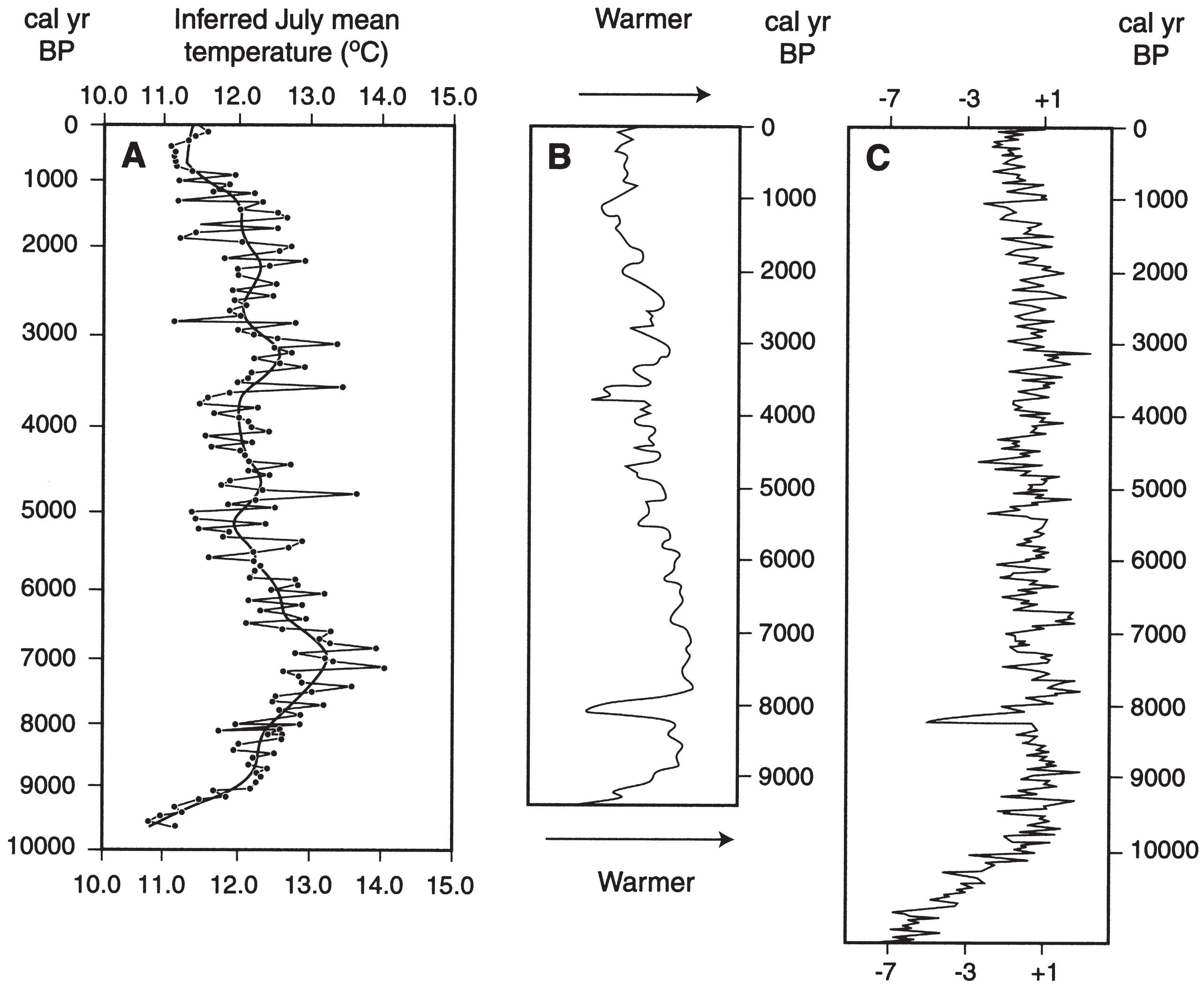

Figure 7 Comparisons of the (A) inferred July mean temperature reconstruction at Tsuolbmajavri (from Figure 6) with (B) the relative x-ray density of proglacial sediments at Vuolep Allakasjaure, northern Sweden (from Karlén, 1998) and (C) the temperature reconstruction from GISP2 $\delta^{18}O$ isotope record from Greenland (from Alley *et al*., 1997), calculated as deviation from the average values during the *c*. 2000 yr prior to the 'Little Ice Age'.

timescales. Palaeolimnological proxies such as chironomids and diatoms are currently being studied at Tsuolbmajavri (Korhola *et al*., 2000) These organisms have the potential to detect finer-scale climate changes.

Cool, moist early Holocene 9900–8200 cal. yr BP

The reconstructed climate of the earliest Holocene indicates fairly low but steadily rising summer temperatures. T_{jul} at *c*. 9500 cal. yr BP is comparable with present values, whereas at *c*. 9300 cal. yr BP it is already above 12.0°C. The rise of T_{jul} may reflect the melting of the Scandinavian ice sheet. According to the deglaciation model of Lundqvist (1991), the ice sheet at 9400 ^{14}C yr BP (*c*. 10 500 cal. yr BP) still covered much of northern Sweden and reached to western Finland. Such a large ice sheet would typically generate pronounced anticyclonic conditions which would induce a cold climate with strong katabatic winds to the periglacial area (e.g., Putnins, 1970). The ice margin then rapidly retreated and the ice sheet was probably completely melted at 9000–8500 ^{14}C yr BP (*c*. 10 000–9500 cal. yr BP) (Lundqvist, 1991).

However, a factor that must be taken into account in all pollen-based climate reconstructions is the possible timelag in vegetation's response to rapidly changing climate (Birks, 1981). This is particularly relevant in connection with the early-Holocene climate warming, since the Late Weichselian distribution ranges of the main tree species may have been located hundreds of kilometres away from northern Fennoscandia (e.g., Huntley and Birks, 1983). A lag in tree migration would produce an apparent gradual warming trend in pollen-based reconstructions although the real climate change might have been sudden (Birks, 1981). The only way to assess this issue is to compare the pollen-reconstructed climate with other independent records. Sediment evidence from the proglacial lake Vuolep Allakasjaure in northern Sweden, *c*. 250 km southwest of Tsuolbmajavri, indicates a gradual warming from deglaciation at 8500 ^{14}C yr BP to 7800 ^{14}C yr BP (*c*. 9500–8500 cal. yr BP) (Karlén, 1998), a similar pattern to the Tsuolbmajavri T_{jul} reconstructions (Figure 7). A similar rising trend at 9800–8500 cal. yr BP is also suggested by a temperature curve calibrated from speleothem $\delta^{18}O$ record from Søylegrotta cave in northern Norway (Lauritzen and Lundberg, 1999).

The direct temperature measurements from the borehole of the GRIP core in Greenland indicate a gradual warming from 11 500 cal. yr BP onwards and that the maximum Holocene temperatures were not reached before 8000 cal. yr BP (Dahl-Jensen *et al*., 1998). Thus, gradual warming may have been a widespread pattern in northern Europe and is not an artifact due to the gradual lag response of vegetation or to local deglaciation history. On the other hand, the rise of $\delta^{18}O$ values in the GISP2 and GRIP cores to the maximum Holocene levels had occurred already by 10 000–9000 cal. yr BP (Dansgaard *et al*., 1993; Stuiver *et al*., 1995), i.e., about 2000 calibrated years earlier than the Holocene maximum T_{jul} values at Tsuolbmajavri.

Reconstructed P_{ann} during the early Holocene is also low at first but rises rapidly to its Holocene maximum values between *c*. 9550 and 7000 cal. yr BP. As T_{jul} is comparatively low, the reconstructed climatic pattern at *c*. 9550 and 8200 cal. yr BP corresponds with the present climate of the Norwegian north coast and suggests a consequently more oceanic climate in northwestern Finland than at present. Similar conclusions for an early-Holocene enhanced oceanicity have been made on the basis of pollen stratigraphy in northern Sweden by Berglund *et al.* (1996) and in the northernmost parts of Finland and Norway by Seppä (1996). There is also evidence to suggest that an oceanic climate was dominant not only in northern Fennoscandia but also more extensively in northwestern Europe. In southwest Norway, for example, the early-Holocene high pine tree-line and advanced glacier positions, as reflected by proglacial lake sediments, suggest that warm summers were compensated by high winter precipitation from 9000 ^{14}C yr BP to 7300 ^{14}C yr BP (*c*. 10000–8000 cal. yr BP) (Dahl and Nesje, 1996), and the available lake-level evidence indicates a generally moist climate in western Europe at 9000 ^{14}C yr BP (*c*. 10000 cal. yr BP) (Yu and Harrison, 1995).

Mid-Holocene warm period *c*. 8200–5700 cal. yr BP

Towards the mid-Holocene the summer temperatures at Tsuolbmajavri rose gradually and reached their Holocene maxima between *c*. 7950 and 6750 cal. yr BP, when reconstructed T_{jul} was 1.7–1.6°C higher than at present and 1.9–1.8°C higher than at *c*. 700 and 300 cal. yr BP. P_{ann} started to decrease rapidly at *c*. 7000 cal. yr BP, i.e., during the time of highest T_{jul}.

There are much data to support the timing of the reconstructed warmest period from northern Fennoscandia. Proglacial lake-sediment records indicate that most of the studied Fennoscandian glaciers had disappeared by *c*. 8000 ^{14}C yr BP (*c*. 8800 cal. yr BP) and that there were only sporadic phases of advancing glaciers from *c*. 8000 to *c*. 6000 ^{14}C yr BP (*c*. 8800–6800 cal. yr BP) (Nesje *et al.*, 1994). The data of Karlén (1998) from Vuolep Allakasjaure is in very close agreement with the reconstructed T_{jul} from Tsuolbmajavri (Figure 7). These data, based on x-ray analysis of a sediment core, indicate that the glacier was smallest from 7200 to 6100 ^{14}C yr BP (*c*. 7950–6950 cal. yr BP). The close correspondence between the results from Tsuolbmajavri and Vuolep Allakasjaure not only strongly suggests that the warmest period in northern Fennoscandia was from *c*. 7950–6750 cal. yr BP but also reflects the importance of ablation-period climatic conditions in controlling glacier mass balance.

The pine megafossil evidence from northern Finland, including the Tsuolbmajavri area, also suggests a clear climate stage during which the pine tree-line was located at *c*. 200 m higher altitude than at present. It also indicates a later beginning and a later ending for the period. The maximal occurrence of pine megafossils from above and beyond the present pine tree-line in Finland dates to 5200–3800 ^{14}C yr BP (*c*. 5950–4150 cal. yr BP) (Eronen and Zetterberg, 1996). The chronological difference between the pollen-inferred warmest period and the maximum occurrence of pine megafossils may indicate that the expansion of pine to high altitudes was amplified by decreasing precipitation from *c*. 7000 cal. yr BP onwards. Of the north-Fennoscandian tree species, pine is best adapted to dry, nutrient-poor sites and favoured by thin snow beds (Kullman, 1986; 1992; Willis *et al*., 1998). Alternatively, it is possible that the pine megafossil data are somewhat biased due to poor preservation of megafossils during the warm and dry periods of the Holocene since the preservation of the pines in northern Fennoscandia indicates that they were often inundated by rising lakewater levels (Eronen and Zetterberg, 1996). However, a case can also be made that the asynchronity is an artifact, caused by the tendency to obtain radiocarbon dates that are too old from lake-sediment bulk datings (see Oldfield *et al*., 1997; Barnekow *et al*., 1998).

Combined with the high summer temperatures, the low P_{ann} (below 450 mm yr^{-1} between *c*. 6350 and 6200 cal. yr BP) must have had strong hydrological and ecological consequences. Evidence for this is provided by diatom and cladoceran stratigraphies from two lakes in hydrologically sensitive fluvioglacial esker terrain *c*. 100 km southeast from Tsuolbmajavri. Qualitative reconstructions indicate that lake levels were probably 4–5 m below the present level during the driest period at 6000–5000 ^{14}C yr BP (6800–5700 cal. yr BP) (Hyvärinen and Alhonen, 1994; Eronen *et al*., 1999).

Late-Holocene cool, moist climate *c*. 5700 cal. yr BP to present

The mid-Holocene cooling at *c*. 5700 cal. yr BP has been detected by different methods from various parts of northern Fennoscandia. Pollen evidence suggests that the retraction of the pine tree-line in northernmost Norway started about at this time (e.g., Hyvärinen, 1975; Seppä, 1996) and proglacial lake evidence suggests activation of glaciers in Sarek, northern Sweden (Karlén, 1988). After the cooling at *c*. 5700 cal. yr BP, T_{jul} remained steady until *c*. 3950 cal. yr BP. Interestingly, our results indicate a warm and dry period at *c*. 3600–3200 cal. yr BP. There is no previous pollen-stratigraphical or lake-level evidence for such a climatic period from northern Finland. However, as can be seen in Figure 7, this period is roughly synchronous with indications of a warmer climate in the Vuolep Allakasjaure proglacial lake-sediment record (Karlén, 1998) and higher $\delta^{18}O$ values at 3500–3300 cal. yr BP in the GISP2 isotope record from Greenland (Alley *et al*., 1997).

Since *c*. 3200 cal. yr BP, there has been a gradually decreasing trend in reconstructed T_{jul}. This late-Holocene cooling is documented also by the activation of a number of Fennoscandian glaciers from 4000–3000 ^{14}C yr BP (4400–3200 cal. yr BP) onwards (Nesje *et al*., 1994; Snowball and Sandgren, 1996), suggesting that the rise in P_{ann} is partly due to an increase in winter snowfall (see Dahl and Nesje, 1996). All three data sets in Figure 7 show consistent cooling trends during the last *c*. 1500 years. It can therefore can be proposed that that the second millennium AD was the coldest in northern Europe since deglaciation.

A consequence of decreased late-Holocene summer temperatures and a gradual increase of P_{ann} in northern Fennoscandia would be an increasing effective humidity and soil moisture. Aario (1943) was the first to obtain evidence for such a trend in northern Fennoscandia. His peat- and pollen-stratigraphical records from Petsamo, northwest Russia, show a distinct activation of paludification processes at *c*. 4500–4000 ^{14}C yr BP (5100–4400 cal. yr BP) and subsequent spread of peatlands. This increased distribution of peatlands is also reflected by the rising *Sphagnum* spore values in many pollen diagrams in northern Fennoscandia (e.g., Hyvärinen, 1975; Seppä, 1996), as well as in increasing relative humic-acid values (Reinikainen and Hyvärinen, 1997) and reconstructed total lakewater organic carbon content in lake-sediment records (Seppä and Weckström, 1999).

Oceanic and atmospheric circulation dynamics

The early-Holocene reconstructed oceanic climate suggests a different atmospheric circulation pattern than at present when the oceanic climate is predominantly confined to the western side of the Scandes Mountains. In the light of the present broad-scale climatic setting in north Europe, such a climatic picture is probably related to the intensification of the western air flow and associated penetration of the rain-bringing cyclones to Fennoscandia. The weather in Fennoscandia, under a westerly type of circulation, differs greatly from one region to another, depending on the intensity and the trajectories of the migrating cyclones (Johannessen, 1970). At present, the westerly air flow and the cyclonic activity over Fennoscandia are most intense in winter when the sea surface west of Fennoscandia is warmer than the air masses blowing over it and the evaporation from the sea is comparatively great. Hence, the westerly circulation generally brings in mild, moisture-bearing air masses and creates large temperature anomalies over Fennoscandia (Johannessen, 1970). In summer the sea surface of the North Atlantic is colder than that of the continent to the east and the Atlantic air masses entering Fennoscandia in summer are relatively cool and dry. The air temperatures and summer precipitation are relatively low on the western side of the Scandes Mountains whereas on the leeward side high temperatures and convective precipitation usually prevail, even during enhanced westerly circulation. Thus, we may infer that high P_{ann} was, to a great extent, as a result of increased winter precipitation.

This interpretation of a more intense early-Holocene westerly air flow, based on the reconstructed climate parameters, is consistent with elevated $\delta^{18}O$ values in calcareous lacustrine sediment in a hard-water lake in Abisko, northern Sweden (Hammarlund and Edwards, 1998). The relative enrichment of the $\delta^{18}O$ suggests an enhanced moisture transport from the Atlantic and, consequently, a pronounced westerly air flow over the Scandes to the presently dry lee-side (Berglund *et al*., 1996; Seppä and Hammarlund, 2000). The $\delta^{18}O$ values decrease gradually from 8500 to 6500 ^{14}C yr BP (*c*. 9500–7400 cal. yr BP), consistent with the general decrease in reconstructed P_{ann} at Tsuolbmajavri. These features may reflect the attenuation of the western air flow over northern Fennoscandia.

Support for the reconstructed early-Holocene climate patterns and a strong westerly air flow comes also from General Circulation Model simulation results. Simulations by the Community Climate Model show that the early-Holocene insolation patterns induced an enhanced general land-sea temperature and pressure contrast in northern Europe which led to a stronger-than-present Iceland low pressure (Harrison *et al*., 1991; 1992) while the subpolar low-pressure belt was shifted northward. Such a climatic setting would have resulted in intense zonal air flow from the North Atlantic which gave rise to higher winter temperatures and P_{ann} in northern Europe (Harrison *et al*., 1991; Huntley and Prentice, 1993).

The broad-scale, spatially coherent shift from the cooler and moister early Holocene to a warm and increasingly dry mid-Holocene suggests a major change in atmospheric circulation. The most obvious explanation is a change from the dominance of the westerly air flow to a more meridional flow pattern. Although the current circulation pattern over the northeastern Atlantic and northern Fennoscandia is dominantly zonal (Figure 8), it may often vary from this mode, and often for a considerable period of time the middle-latitude jet-stream is blocked by a deep and warm anticyclone in the pressure system of the upper air and at the earth's surface (Johannessen, 1970). Blocking anticyclones are associated with low values of the zonal index and are produced by well-established, quasi-stationary positive pressure anomalies in the wave pattern of the westerly air flow (Davies *et al*., 1997). In such circumstances the westerly air flow over the North Atlantic and Fennoscandia is replaced by a meridional or quasi-meridional circulation (Johannessen, 1970). Depending on the position and character of the high-pressure system, this circulation type often produces a southerly air flow and generates high summer temperatures and long dry spells (Figure 8).

Earlier, Yu and Harrison (1995) and Harrison *et al.* (1996) have interpreted that the low lake levels in northern Europe from 8000 to 5000 ^{14}C yr BP (*c*. 8800–5700 cal. yr BP) resulted from the increased persistence of blocking anticyclones over Fennoscandia. They argue that such a climatic pattern could have been produced by the disappearance of the Scandinavian ice sheet between 8500 and 7500 ^{14}C yr BP (*c*. 9500–8250 cal. yr BP) (Lundqvist, 1991) and the higher-than-present summer insolation and/or by the larger-than-present area covered by the Baltic Sea.

It is significant that marine evidence from the western entrance of the Barents Sea, a shallow continental shelf sea north and northeast of the Norwegian coast (Figure 1), supports the proposed early- to mid-Holocene change in dominant atmospheric circulation type. The deep-sea sediment evidence indicates lower SST from 8000 to 5000 ^{14}C yr BP (*c*. 8800–5700 cal. yr BP) than during the early or late Holocene (Sarnthein *et al*., 1995; Hald and Aspeli, 1997). Interestingly, in the Barents Sea climate model the high Barents Sea SSTs are connected to low air pressure, intensification of cyclonic circulation, reduced sea-ice cover and increased influx of Atlantic waters to the Barents Sea. Low SSTs, in turn, occur during higher air pressure and decreased inflow of the Atlantic water (Ådlansvik and Loeng, 1991). These features are consistent with the reconstructed climatic and atmospheric circulation patterns in northern Fennoscandia and suggest closely coupled atmosphere-ocean interactions in north Europe during the Holocene.

Late-Holocene climate with low temperatures and increasing annual precipitation suggests reduced influence of the blocking anticyclones, probably due to the gradual decrease of the summer solar insolation. A more boreal type of climate with cool summers is reflected by the increase of *Sphagnum*-dominated peatlands and especially by the immigration and expansion of spruce. The current circulation pattern with a dominant zonal flow, alternating high-pressure systems and associated relatively cool, moist summers seems to have become established during the first and second millennia AD.

Conclusions

The results from this study show that long-term climate changes in northern Fennoscandia can be reconstructed quantitatively, with reconstruction errors of *c*. 1.0°C (July mean temperature) and 340 mm (annual precipitation), from pollen-stratigraphical data. The resulting reconstructions are generally consistent with other independent proxy records from northern Fennoscandia, although the quality of the chronology of most of the records is a limiting factor in detailed correlation and comparison between proxies from different areas.

The reconstructions suggest that the early-Holocene (9900 to 8200 cal. yr BP) climate in northwest Finland was more oceanic than at present, with low (*c*. 11.0°C) but rapidly increasing July mean temperatures and with annual precipitation ranging from 600 to 800 mm. From the early-Holocene oceanic climate, July mean temperatures increased to the highest Holocene values at 7950 cal. yr BP, when they rose above 12.5°C and stayed high to 6700 cal. yr BP. The maximum July mean temperatures at *c*. 7300 cal. yr BP were *c*. 13.0°C which is 1.6–1.7°C higher than at present.

The cooling of July mean temperatures had started already at *c*. 6750 cal. yr BP. A clear cooling trend characterized the last two millennia and the second millennium AD has been the coolest since the early Holocene with reconstructed July mean temperatures of 11.0–11.6°C. Parallel with the late-Holocene cooling has been the increase in precipitation. Together these have led to an increased effective moisture, which is reflected elsewhere in northern Fennoscandia by the rise of lake levels (Hyvärinen and Alhonen, 1994) and by the expansion of peatlands (Hyvärinen, 1975; Seppä, 1996).

Our data also indicate that superimposed on the general climatic trends there may have been higher-frequency climate excursions in northwest Finland. In the light of our present data it is not possible to state if the reconstructed warm and cold intervals (e.g., at *c*. 3600–3200 cal. yr BP) are 'real' and reproducible and if they are linked to the suggested cyclic climate changes in the North Atlantic (Bond *et al*., 1997; Chapman *et al*., 2000) and suggested abrupt climate changes on the Fennoscandian mainland (Karlén,

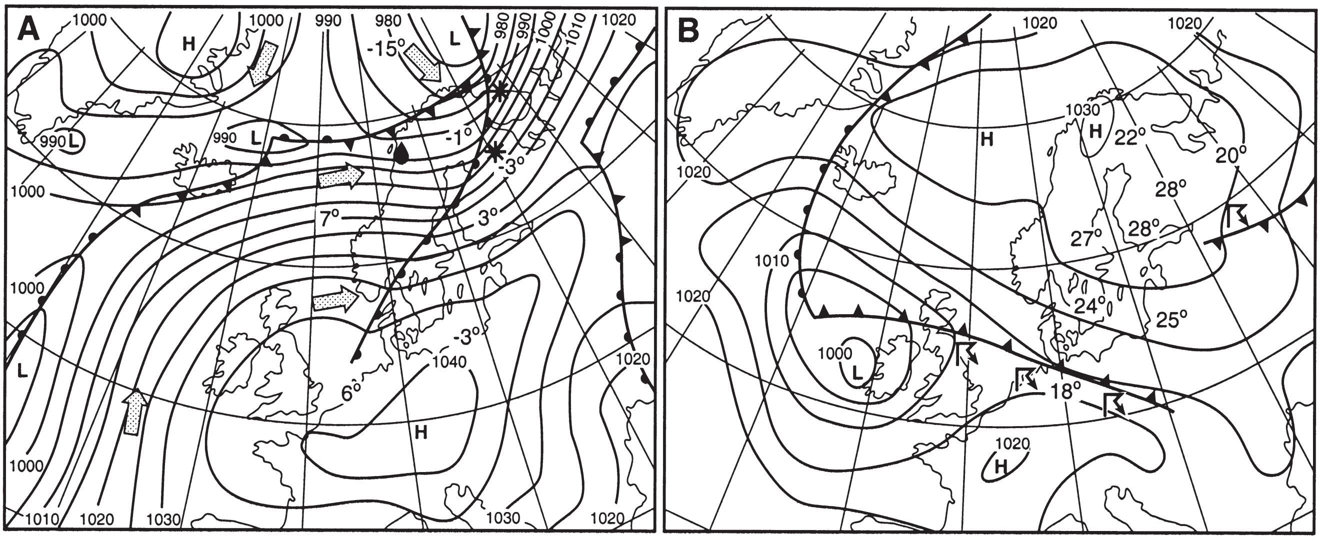

Figure 8 (A) An assumed dominant atmospheric circulation pattern for the earliest Holocene (to *c.* 8200 cal. yr BP) in northern Fennoscandia with enhanced westerly air flow from the North Atlantic and low air-pressure cell on the Barents Sea (27 January 1981). (B) An assumed dominant atmospheric circulation pattern for the early to mid-Holocene (*c.* 8200–5700 cal. yr BP) with anticyclonic summer conditions over northern Fennoscandia (30 July 1980) (redrawn from the *Atlas of Finland*, 1987).

1998; Nesje *et al.*, 2000; 2001). The occurrence of these excursions must be tested by additional well-dated high-resolution climate proxy cores from critical regions in northern Fennoscandia.

The roughly similar reconstructed past climatic patterns between the Greenland ice-core data (Dahl-Jensen *et al.*, 1998), marine core data from the North Iceland shelf (Eiriksson *et al.*, 2000), and our data indicate that a linked North-Atlantic ocean-atmosphere circulation system has been the predominant driving force for the Fennoscandian mainland climate during the Holocene. The results suggest that long-term solar insolation patterns did not directly govern the climate. This is especially so in the case of the reconstructed early-Holocene oceanic climate as insolation patterns would suggest warm summers (8% higher July solar insolation) and cold winters (8% lower January solar insolation), i.e., a more continental climate, at *c.* 9000 ^{14}C yr BP (*c.* 10000 cal. yr BP) (Kutzbach and Ruddiman, 1993).

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