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Northern high-latitude climate change between the mid and late Holocene – Part 1: Proxy data evidence

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Abstract

In this paper we try to develop a quantitative understanding of the absolute change in climate between the mid-Holocene ~6000 yr BP (6 ka) and the preindustrial period ~1750 AD (0 ka) in the northern high latitudes. This has been performed using available quantitative reconstructions of temperature and precipitation from proxy data. The main reason for comparing these two periods is that the summer insolation in the northern high latitudes was higher at 6 ka than 0 ka due to orbital forcing. Another reason is that it gives us the opportunity to quantitatively compare results from proxy data with results from several climate model simulations for the same periods by using data from the Palaeoclimate Modelling Intercomparison Project. Another aim has been to try and quantify the uncertainties in the proxy data reconstructions. The reconstructions indicate that the northern high latitudes were 0.96±0.42°C warmer in summer, 1.71±1.70°C warmer in winter and 2.02±0.72°C warmer in the annual mean temperature at 6 ka compared to 0 ka. The warmer climate in summer around 6 ka BP was most likely directly related to the higher summer insolation whereas the warmer climate in annual mean and winter temperature may possibly be explained by internal physical mechanisms such as heat stored in the oceans during summer and released during the cold season or by changes in the vegetation causing albedo changes that may affect seasonal temperatures differentially. For the future there is a great need to reduce the errors of the predictions as well as improving our understanding of how a proxys respond to changes in environmental variables.

1 Introduction

During recent decades the northern high latitudes have experienced significant warming, which is larger than elsewhere on globe (e.g. Moritz et al., 2002; Brohan et al., 2006). Averaged Arctic temperatures have increased at almost twice the global average rate in the past 100 yr. Observations since 1961 show that temperatures have risen by more than 2°C in Arctic areas (IPCC, 2007). The observational evidences are generally consistent with climate model simulations that include increased greenhouse gas concentrations and other observed external forcings (Holland and Bitz, 2003) but uncertainties in regional climate predictions, which are relying on our current understanding of climate-influencing processes in the various components of the climate system still exists (IPCC, 2007). A better understanding of the regional climate evolution is crucial for understanding present climate dynamics and a prerequisite to meet expressed needs of improved climate forecasting capabilities. This necessitates a longer time perspective that reaches beyond the information available from instrumental records (Jones et al., 2009).

Quantitative reconstructions of climate have been produced by using different types of proxy data from natural archives, like biological proxies preserved in lake and marine sediments (e.g. pollen, diatoms, chironomids or foraminifera), the width of tree-rings or the oxygen isotope composition in ice, speleothems or lake sediments. Several of the reconstructions from the northern high latitude regions reveal the Holocene epoch as, although having a much more stable climate than the preceding Weichselian glaciations, showing a general long term cooling during the last 6000 yr. The temperature proxy records, mainly based on pollen (Seppä and Birks, 2001; Heikkila and Seppä, 2003), broadly mirror the simultaneous trend of reduced incoming solar radiation in the boreal summer. The main reason for this cooling trend is orbital factors (tilt and precession) that produced a declining summer insolation in the northern hemisphere (Berger and Loutre, 1991). While the solar insolation in boreal summer was maximal at around 11 ka BP, the temperature maxima in the proxy records are in some areas delayed up to 4000 yr as a result of the cooling effects of melting of the big continental ice sheets (Kaufmann et al., 2004). From pollen studies, Davies et al. (2003) suggest that the so called Holocene Thermal Maximum (HTM) occurred across a wide area of northern Europe at around 6 ka BP. The long-term temperature trend caused by the change in orbital forcing was amplified by several strong positive feedbacks, i.e. the ice-albedo feedback, the sea-ice insolation feedback, and the tundra-taiga feedback.

A further mechanism of importance is the slow release of the excess heat stored in the oceans due to their large heat capacity (Renssen et al., 2006).

The aim of this paper is to develop an improved quantitative understanding of the *absolute change in climate* between the mid-Holocene ∼6000 yr BP (6 ka) and the preindustrial period ∼1750 AD (0 ka) in the northern high latitudes, using available quantitative reconstructions of temperature and precipitation from proxy data. Compared to other regions a large number of quantitative reconstructions are available for mid-Holocene over the northern high latitudes. A main reason for choosing the two selected periods is that it gives us the opportunity to quantitatively compare results from proxy data with results from several climate model simulations for the same periods by using data from the Palaeoclimate Modelling Intercomparison Project (PMIP, http://pmip.lsce.ipsl.fr/ and http://pmip2.lsce.ipsl.fr/). Such a proxy-data vs. model-data comparison is undertaken in a companion paper (Zhang et al., 2009). An important reason why PMIP chose to undertake simulations for the periods 6 ka and 0 ka is that summer insolation in the northern hemisphere was still high whereas most of the big ice sheets (e.g. North America) were gone at 6 ka, and thus the climatic effect from the former big ice sheets was small.

A direct quantitative model-data comparison requires that the proxy data are associated with realistic estimates of the uncertainty in the reconstructed climate variables. Therefore, a central issue in this paper is to quantify these uncertainties and this leads us to discuss the reconstructed changes in light of the estimated uncertainties. The subsequent model-data comparisons in the companion paper will then help in clarifying inconsistencies and ambiguities in the climate change as recorded by the proxies as well as the physical mechanisms behind the changes as recorded in the model simulations, and so we can further evaluate the role of orbital forcing as well as feedback processes to climate change.

2 Material and methods

2.1 Available climate reconstructions based on proxy data

Through a screening of published quantitative temperature and precipitation reconstructions based on proxy data from north of ~60° N covering the mid-Holocene (6000 yr BP) and the pre-industrial (~1750 AD) periods, a total number of 124 reconstructions from 71 different sites were found (Fig. 1, Table 1). The digital data sets were obtained from the NOAA data base (http://www.ncdc.noaa.gov/paleo/recons.html) or directly from respective author. The screening illustrates that the information in the proxy records have a large over-representation towards summer temperatures (79 out of 124), whereas only 17 represent annual mean temperature, 7 winter (Jan) temperature,18 annual mean precipitation and 6 winter precipitation (according to the original interpretations done by the various investigators; see references in Table 1). The geographical distribution of the records is also not uniform; there is a large bias towards the land areas surrounding the North Atlantic sector, especially Fennoscandia (61 out of 124).

A major part of the reconstructions are terrestrial reconstructions (113 out of 124) and most of these are reconstructed from biological proxies, like pollen, chironomids and diatoms, where the abundance of various species is calibrated using transfer functions that are determined from the distributions of species in modern surface lake or ocean hydrography over a wide geographical range, using the assumption that recent variations in species between different climatic regions are the same as variations caused by changes in climate over time at one and the same site. Other terrestrial proxies used are tree-ring width (1), $\delta^{18}O$ in speleothems (1) and ice core (1), borehole temperature (2) and density of sediment in combination with pollen (5). Reconstructions of sea surface temperatures (SSTs) are based on proxies that make use of the chemical composition (alkenone unsaturation index, $\delta^{18}O$ of foraminiferal shells) or abundance of planktonic organisms (diatoms and foraminifera). For detailed information on how each proxy type was calibrated to either temperature or precipitation, see references in Table 1.

2.2 Methods

For each collected proxy record, a 100 yr mean value of the reconstruction at 6 ka and 0 ka was estimated. These mean values were calculated to correspond as closely as practically possible to the mean values for the periods 6000 yr BP ±50 yr and 1750 AD±50 yr, respectively. The main reason for averaging over these 100 yr time windows is to be able to compare the proxy values directly with the output from the PMIP climate model simulations (Zhang et al., 2009). To estimate the climate change between the two periods, the value at 0 ka was subtracted from the value at 6 ka. For records that were not available digitally (47 out of 124), a value was read visually from the graph for each time slice. For records with uncalibrated radiocarbon ages, the radiocarbon dates were calibrated into years before present using the program OXCAL 4.0 (Bonk Ramsey 2008). For each value of climate change that was calculated a value of uncertainty was estimated, using the approach described below.

2.2.1 Estimating a 100 yr mean value

For each time slice (6 ka and 0 ka) a weighted average of data for the climate variable during the 100 yr was calculated. This was done by first calculating a "preliminary" weight, v_i for each value of the reconstruction (X_i) (either temperature or precipitation) that contributes to the time slice average, $\overline{X}$. The weights (Eq. 1) were introduced to ensure that the observations closer to the midpoint of the time slice got a higher weight. The final weights (w_i) where then calculated by dividing with the sum of all the preliminary weights (v_i) so that the sum of all weights add up to one (Eq. 2).

$$v_i = e^{-|t-t_i|/100} \tag{1}$$

$$w_i = \frac{v_i}{\sum_{i=1}^{N} v_i} \tag{2}$$

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Here the time is measured in yr and t_i is the mid-year of the time-slice. The weighted average at a certain time slice (6 ka or 0 ka) is then;

$$\overline{X} = \sum_{i=1}^{N} X_i \times w_i \tag{3}$$

Finally, the change in the climate variable between the two periods is calculated as;

$$\Delta\overline{X} = \overline{X}_{6\,\mathrm{ka}} - \overline{X}_{0\,\mathrm{ka}} \tag{4}$$

2.2.2 Estimation of uncertainty

For each calculated parameter (temperature or precipitation) of climate change, between 6 ka and 0 ka mainly three types of uncertainties exist.

1. Calibration uncertainty, σ_c^2,
2. Dating uncertainty, σ_d^2,
3. Reading (from graph) uncertainty, σ_r^2.

It is assumed that each uncertainty is a normally distributed stochastic variable and that the three errors are independent of each other. The three errors all have the different variances; σ_c^2, σ_d^2, σ_r^2.

$$\sigma_\mathrm{tot}^2 = \sigma_\mathrm{c}^2 + \sigma_\mathrm{d}^2 + \sigma_\mathrm{r}^2 \tag{5}$$

An approximate 95% confidence interval of the true uncertainty for the reconstructed measurement X_i is then $2\sigma_\mathrm{tot}$; Below, we describe how we estimate each of the three errors, and how they are combined to give estimates of the uncertainty in the "reconstructed" climate change between 6 ka and 0 ka, i.e. the uncertainty in $\Delta\overline{X}$.

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Calibration uncertainty, σ_c

The calibration uncertainty is due to the incapability of the proxy data to perfectly portray past variations of the climate variable of interest. This uncertainty is usually given by the original investigators in their text as a root mean square error of prediction (RMSEP) or a sample specific error of prediction (SSEP) of the reconstructions. The variance of the "total" uncertainty in the estimates of $\Delta\overline{X}$ due to the calibration uncertainty in both time slices is obtained by adding the variances of the calibration at each time slice.

$$\sigma^2_{\mathrm{c\,tot}} = \sigma^2_{\mathrm{c\,6\,ka}} + \sigma^2_{\mathrm{c\,0\,ka}} \tag{6}$$

If there is more than 1 observation per estimation, however, i.e. if the resolution of the record is higher than 100 yr then the calibration uncertainty has to be adapted (reduced) to account for the number of observations (N) in the time slice. Furthermore, since we want the observations that lie closer to the midpoint of time slice to have larger impact, we use the weights from Eq. (2) to estimate the total uncertainty in $\Delta\overline{X}$ that is due to the calibration uncertainty.

$$\sigma^2_{\mathrm{c\,tot}} = \sigma^2_{\mathrm{c\,6\,ka}} \sum_{i=1}^{N_{6\,\mathrm{ka}}} w^2_{i\,6\,\mathrm{ka}} + \sigma^2_{\mathrm{c\,0\,ka}} \sum_{i=1}^{N_{0\,\mathrm{ka}}} w^2_{i\,0\,\mathrm{ka}} \tag{7}$$

Dating uncertainty, σ_d

With "dating uncertainty", we here mean the uncertainty in the climate variable in question that is due to the uncertainty in the age-depth model in the relevant sedimentary archive. Although a dating uncertainty is originally (and obviously) given in units of time, we need to find a way for how this uncertainty can be transformed into units of the climate variable.

The "real" dating uncertainty, i.e. in time units at 6 ka and 0 ka, were first calculated by interpolation between the errors of nearby dating points. Then the minimum and maximum value of the climate estimate within each time interval (I) was picked to estimate the standard deviation of the dating uncertainty given in °C or mm rain.

$$\sigma_{\mathrm{d},I} = \frac{x_{\max} - x_{\min}}{2} \tag{8}$$

Finally, the variance of the "total" dating uncertainty for the quantity $\Delta\overline{X}$ was calculated by adding the variances of the uncertainties at 6 ka and 0 ka together.

$$\sigma^2_{\mathrm{d\,tot}} = \sigma^2_{\mathrm{d\,6\,ka}} + \sigma^2_{\mathrm{d\,0\,ka}} \tag{9}$$

Reading (from graph) uncertainty, σ_{r}

In the 47 cases when data were not available digitally, the reading uncertainty was estimated to 0.5 mm per reading in a printout of the original graph at an approximate size of $10\times20\,\mathrm{cm}^2$. This uncertainty in reading is assumed to correspond directly to one standard deviation of the climate variable. The variance of the "total" reading uncertainty for $\Delta\overline{X}$;

$$\sigma^2_{\mathrm{r\,tot}} = \sigma^2_{\mathrm{r\,6\,ka}} + \sigma^2_{\mathrm{r\,0\,ka}} \tag{10}$$

So, if the scale of a diagram is Y units mm^{-1} and assume that then;

$$\sigma^2_{\mathrm{r\,tot}} = \frac{Y}{\sqrt{2}} \tag{11}$$

Averaging uncertainties

In the following discussion, we are sometimes interested in presenting results averaged over several proxy series (either within the same site or within a larger region). In these cases, we need to find estimates of the total uncertainty in the averaged climate data. To achieve this, the variance of the total uncertainty in the averaged data is calculated by dividing the sum of the variances of the total uncertainties for each proxy by the number of proxy series (N_p).

$$\overline{\sigma} = \frac{1}{N_p}\sqrt{\sum_{i=1}^{N_p} \sigma_i^2} \tag{12}$$

3 Results and discussion

3.1 Climate change between 6 ka and 0 ka

A large majority of the reconstructions (107/124) indicates that temperatures were on average higher at 6 ka than during the recent preindustrial era (Fig. 2). According to an unweighted average of $\Delta\overline{T}$ across all seasonally separated proxy series, the climate of the northern high latitudes at 6 ka was on average 0.96±0.42°C warmer in summer and 2.02±0.72°C in the annual mean, in comparison to 0 ka. Note, however, that the GISP2 annual mean temperature reconstruction from Greenland (Alley et al., 2000) is excluded from this quantitative analysis due to lack of information about its uncertainty. For winter data the average change is a cooling from 6 ka to 0 ka by 1.71±1.70°C, but as the uncertainty is about as large as the average change one cannot conclude that the average winter cooling is statistically significant. It is noteworthy that the estimated change in the annual mean temperature is larger than in both winter and summer. There may be several possible reasons for this behavior (see Sect. 3.3).

Only for the summer season the proxy records are sufficiently numerous to allow a quantitative comparison of temperature changes in different regions. The estimated changes in summer temperature show regional differences (Fig. 3) with Siberia having seen the most pronounced cooling of 2.24±0.88°C. A cooling is also seen for Fennoscandia with 0.94±0.42°C and for the Nordic Seas with 1.04±0.89°C, while the temperature changes over North America and Iceland are within the $2\overline{\sigma}$ uncertainties for the regional averages, 0.53±0.80°C and 0.15±1.26, respectively. The SST reconstructions from foraminifera have not been included in the estimated average for the Nordic Seas. The reason for this is that in contrast to diatoms and alkenone producing algae, which live in the upper 50 m of the water column, the foraminifera are found deeper down near the permanent thermocline. Usually this lower part of the ocean is unaffected by the near-surface warming during the summer season (Jansen et al., 2008). The average temperature change recorded by the foraminifera is −0.64±0.98°C, which in opposite to most of the other records indicate that warmer conditions at 6 ka compared to 0 ka. Given the large uncertainty, however, one must regard the change as statistically insignificant.

Siberia is the only region for which the number of proxy records for winter and annual mean temperatures is large enough for any meaningful comparison between seasons within a region. In addition to the nine summer temperature proxies, there are four proxies each that are interpreted as winter and annual mean temperature records. These latter reconstructions indicate significantly warmer winter temperatures of 2.66±1.46°C as well as annual mean temperatures of 2.39±1.46°C at 6 ka compared to 0 ka. Thus, the average temperature change in the winter and annual mean for the Siberian proxies are about the same as the above reported change for the summer proxies from this region. Taking account for the large uncertainties, it is not possible to judge if the temperature change differed between the seasons.

Solely looking at the orbital forcing, we would expect to see a significant cooling in summer and perhaps a small warming in winter from 6 ka to 0 ka (see the companion paper by Zhang et al. (2009e). In summer we would also expect to see a larger change with increasing latitude. Indeed, the available proxy data support the expected overall cooling in summer, but there is no evidence for any clear latitudinal gradients in summer temperature change, nor is there any evidence for any warming in winter. It is not possible to judge solely from a statistical analysis like the one undertaken here, what the reasons might be for the observed changes in the proxy records, or whether the observed changes are physically plausible or not. A systematic comparison with results obtained from climate model simulations will help to better understand the observed changes. Such an analysis is one of the goals of the companion paper by Zhang et al. (2009).

The number of available reconstructions of annual total precipitation from the northern high latitudes is 21 and they are derived from 15 different sites. The average difference in annual total precipitation between 6 ka and 0 ka indicates an overall decrease by 37 mm. Due to large uncertainties associated with the precipitation reconstructions (±130 mm for the $2\overline{\sigma}$ across all records) it is, however, not possible to judge if any significant average change in precipitation has occurred at all. For southern Norway there exist five reconstructions of winter precipitation, which together indicate that the period around 6 ka received 21% less precipitation in winter compared to at 0 ka. However, the uncertainties for these reconstructions lie around 20% (Jostein Bakke, personal communication, May 2009), so it is not possible to judge if any significant change in winter precipitation has occurred even for this comparatively data-rich region.

3.2 Regional availability of proxy data

In an ideal study of the change in climate over the northern high latitudes we would have data evenly spread around the region. However, as mentioned earlier, this is not the case (Fig. 1). There are several reasons for this. Depending on the environment there are different types of available natural archives and thereby there are regional differences in the types of available proxy data. For example, there are no big ice sheets except for the one in Greenland and there are no recent lakes in the middle of Greenland and hence no access to lacustrine sedimentary proxies from this region but instead only ice cores. Also, a certain proxy can be either a better or worse climate indicator depending on the climatic regime. For example, annual mean temperature is probably a more appropriate climatic variable to be reconstructed from pollen data in southern and central Fennoscandia than July temperature, which is presumably better reconstructed from pollen at more northern sites with a shorter growing season (Seppä et al., 2004). Also when performing research aiming at reconstructing past climates, it is desirable to study archives that are as little as possible disturbed by human influence, so that it is primarily a climate signal that is recorded and not effects of human settlement. This is the main reason why most of the data are derived from remote areas like the mountain chains and Greenland. Another limitation is the availability of training sets for calibration the proxy data into temperature or precipitation; for example no modern chironomid training set is available for Russia (Brooks, 2006).

3.3 What is being reconstructed?

As evidenced by our literature survey and data screening, July temperature is the most common climate variable being reconstructed for the mid to late Holocene epoch. However, what is really being reconstructed in these cases is the temperature of the warmest month. Today, this would for continental sites mean July, but at 6 ka this was more likely to be August (Berger, 1979; Zhang et al., 2009). January temperature, or the temperature of the coldest month, can also be reconstructed from pollen data. This is possible because winter climatic conditions are considered to be important for the distribution and regeneration of many plant species, especially those restricted to the most oceanic parts along the west cost of Fennoscandia (Giesecke et al., 2008).

As already mentioned, the estimated average change in annual mean temperature is larger than the average change of summer and winter temperature. There could be several explanations for this behavior. For example, the temperature change in spring and autumn may have been larger than in both winter and summer. However, there is a difference in the number and the spatial distribution of annual mean temperature reconstructions compared to summer and winter reconstructions, which complicate a comparison of results for the different seasons. For locations under marine influence the larger estimated cooling in winter and the annual mean temperatures, compared to summer, could perhaps be explained by summer heat uptake by the ocean that is released to atmosphere during winter; this mechanism would be even more prevailing if the sea-ice cover was less extensive. Otto et al. (2009) have shown that the atmosphere-ocean feedback is the most important in amplifying the effects of the mid-

Holocene insolation forcing and especially during the autumn. Albedo effects due to changes in vegetation, causing larger seasonal effects in spring or autumn, may also be considered as a reason. Another possible reason for the large estimated temperature change seen in the annual mean temperature reconstructions, that cannot be excluded, is that the actual proxy data (predominantly pollen, but also speleothems, oxygen isotopes in ice and borehole temperature measurements in ice) or the transfer functions used to derive the temperature estimates are not sufficiently accurate to permit realistic estimates of past annual mean temperatures. In particular in the case of pollen data, one may perhaps suspect a "seasonal bias" towards summer temperatures that may have a too large influence on the estimated past annual mean temperatures. It is beyond the scope of this paper to speculate further on this matter, but a more thorough investigation of this problem seems worthwhile.

3.4 Comparison of different proxies

The change in summer temperature reconstructed from different terrestrial proxies is not homogenous. On average, chironomids indicate a change of 0.51±0.57°C, and hence the observed average change is not statistically significant. In contrast, pollen and diatoms show significant cooling by 1.23±0.37°C and 1.03±0.67°C, respectively (Fig. 4). One explanation for the differences between the proxies could be that the region of Siberia, which apparently has seen the largest temperature change, is mostly represented by temperature reconstructions from pollen. On the other hand, it should be noted that the error bars for the pollen, diatom and chironomid estimates all overlap, and hence one should not conclude that there are any statistically significant differences between the temperature changes recorded in the three types of proxies.

Chironomids and diatoms live in the lakes. They are dependent on the actual water temperature of the lake but also on catchment driven fluctuations such as pH, water depth, nutrients and dissolved oxygen which may have had a stronger influence on the fauna during certain parts of the Holocene (Brooks, 2006, diatom ref). It is also a problem to convert the water temperature into air temperature since the water temperature

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could be affected by for example glacier melt water and this could cause underestimates of past air temperatures (Brooks, 2006). For the pollen reconstructions there could be problems with human influence on the vegetation as well as long-distance transported pollen (Birks and Seppä, 2004). From Bjørnfjelltjørn in Norway it has been seen that the chironomid-inferred temperatures consistently underestimate the mean July temperatures when compared to pollen thorough out the early and mid-Holocene. Macrofossils from the same period supports the pollen-inferred estimates (Brooks, 2006).

3.5 Comparison between reconstructions from the same and nearby sites

Fennoscandia is the most data-dense region in our survey. This region includes several reconstructions from the same or nearby sites. There are, however, major deviations between the different reconstructions. In Fig. 5, estimated changes between 6 ka and 0 ka for ten reconstructions of summer temperature from the Abisko area (68.33–68.50° N, 18.07–19.12° E) in northern Sweden are shown. The average change in temperature across all records is a cooling by 0.94±0.75°C. Since this is a small area we would expect the various reconstructions to display similar changes in summer temperature. However, this is not the case. Based on the present knowledge it is difficult to evaluate which reconstruction is the most representative of the area. Five of the estimated $\Delta\overline{T}$ values for individual proxies differ from the overall average value by more than twice the standard deviation for the overall mean ($2\overline{\sigma}$), but for four of these records their individual 2σ sigma error bars overlap the overall ($2\overline{\sigma}$) interval. For the fifth reconstruction, which is the tree-ring width reconstruction, both the central estimate and the error bar lie outside the overal $2\overline{\sigma}$ uncertainty interval. In fact, the $\Delta\overline{T}$ value for the tree-ring reconstruction suggests colder summer temperatures at 6 ka compared to 0 ka. The authors of the reconstruction, Grudd et al. (2002), point out that the tree-ring width reconstruction does not express the full range of millennial time scale temperature variation in the Torneträsk area. However, since this is just a comparison of ten

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proxy series it might not be appropriate solely from statistical grounds to consider this reconstruction as an outlier. Noteworthy, two other proxy records (diatoms and chironomids from Lake 850; Laroque and Bigler, 2004; Bigler et al., 2003) also show $\Delta\overline{T}$ values that suggest colder summer temperatures at 6 ka. The warming seen in the three individual records is not significant if their statistical uncertainty is accounted for.

3.6 More on uncertainties

The total uncertainties (2σ) for the estimated $\Delta\overline{T}$ in the individual reconstructions is found to lie between 6.21 and 0.05°C, with the largest uncertainties seen for winter temperature estimates (6.21 to 2.24°C). The largest contribution to the total uncertainty is the uncertainty in the calibration, while the dating and reading uncertainties generally are much lower (Table 1). The dating uncertainty, however, is generally considerably larger for the marine compared to the terrestrial records.

When reconstructing past climate changes it is assumed that the environmental processes that govern the pattern of e.g. the vegetation or the diatom flora in a lake have been the same for the entire period of the reconstruction (e.g. the Holocene). This is most likely not true.

Also there are pitfalls like far-distance pollen, human influence on vegetation, pollen production and climate, biological interactions, identification and morphological limitations (Seppä and Birks, 2004). Another pitfall that is of importance for our study is the spatiotemporal pattern of the Holocene Thermal Maximum. Chironomid reconstructions of July temperature from Fennoscandia from sites near or beyond the present tree-line indicate that the highest July temperatures occurred before or immediately after 8000 yr ago (Giesecke et al., 2008). Low-land sites are rare, but for one of those maximum July temperature was reached after 7000 yr ago (Antonsson et al., 2006). Temperature reconstructions using pollen from Fennoscandia indicate that the July temperature is highest at 8–7 ka while the annual mean temperature is highest at 6.5–5.5 ka (Seppä et al., 2009). Kaufman et al. (2004) have examined the temporal variation of HTM in the western Arctic. They found that HTM occurred around 11–9 ka in Alaska and NW Canada, around 9 ka in the Canadian Arctic islands, and around 7 ka in Hudson Bay. The delayed warming in northeastern America was associated with the cooling effect of the residual Laurentide Ice Sheet.

There is an urgent need to explore methods how to reduce the uncertainties of the reconstructions. Ways of doing this could be by; increasing the replication of records by having a globally coordinated sampling strategy; (today we are comparing results obtained with very different methods); by performing comparisons with model simulations, and, finally, by comparing directly different archives from the same site/area. Also a better characterization of environments by more in situ studies and monitoring could help in reducing the error of calibration.

The influence of other factors than temperature and precipitation (e.g human, pH) could be averaged out by merging temperature estimates from several sites (e.g. Velle et al., 2005; Bjune et al., 2009). In an attempt to improve the performance of chironomid-temperature estimates, Korhola et al. (2002) used a Bayesian statistical method as an alternative to the traditional weighted averaging partial least squares regression (WA-PLS) method for Lake Toskaljavri in northern Finland. The temperature trends produced by both the models were similar but the WA-PLS method had about 1°C higher sample specific errors.

4 Conclusions

This is, to our knowledge, the first study that takes into account all available quantitative reconstructions of proxy data from the northern high latitudes to estimate the change in temperature and precipitation between time slices in the mid- and late-Holocene, and at the same time attempts to quantify the total statistical uncertainties of the observed changes by using the information provided in the original investigations. By taking a simple arithmetic average, the reconstructions indicate that the northern high latitudes were 0.96±0.42°C warmer in summer, 1.71±0.70°C warmer in winter and 2.02±0.72°C warmer in the annual mean temperature at the mid-Holocene (6 ka) compared to the recent pre-industrial era (1700–1800 AD). The warmer temperatures in summer around 6 ka BP was most likely directly related to the higher summer insolation caused by orbital forcing. The warmer climate in annual mean and winter temperature may possibly be explained by internal physical mechanisms such as heat stored in the oceans during summer and released during the cold season or by changes in the vegetation causing albedo changes that may affect seasonal temperatures differentially. One may, however, not exclude systematic errors or biases in different seasonal temperature estimates which complicate comparison between seasons. A better understanding regarding the reasons for the observed changes can be obtained by systematic quantitative comparisons between the observations seen in proxy data and those seen in climate model simulations. Such comparisons are undertaken in a companion paper by Zhang et al. (2009).

The challenge of producing reliable inferred climate reconstructions for the Holocene are great mainly because the expected temperature and precipitation fluctuations during this period are in magnitude similar to the prediction errors of the inference models. There are sometimes large discrepancies between different reconstructions from the same area. The main reason for this may be that the influence of other environmental variables than climate on the proxy is sometimes stronger. A solution to this could possibly be to identify and analyze e.g. a lake that has not been exposed to major fluctuations in variables other than climate or merge together several reconstructions from the same area. For the future there is a great need to reduce the errors of the predictions, to produce training sets for Russia (and other relevant regions) since there is a lack of data from this region, and improve our understanding of how a proxy respond to changes in environmental variables.

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Table 1. (a) Temperature reconstructions from proxy data north of 58° N discussed in this paper.

Site	Lat N	Lon E	Proxy[a]	Variable[b]	6 ka–0 ka	σ_c	σ_d	σ_r	$2\sigma_{tot}$	Reference
Fennoscandia										
Dalmutladdo	69.17	20.72	P	T_{Jul}	1.36	1.08	0.15	0	2.18	Bjune et al., 2004
Barheivatn	69.70	19.85	P	T_{Jul}	2.30	1.08	?	0	?	Bjune et al., 2004
Toskaljavri	69.20	21.47	P	T_{Jul}	0.69	0.73	0.43	0	1.69	Seppä and Birks, 2002
Toskaljavri	69.20	21.47	C	T_{Jul}	0.69	1.00	0.15	0	2.02	Seppä et al., 2002
KP-2	68.80	35.32	P	T_{Jul}	−0.52	0.90	0.28	0	1.88	Seppä et al., 2008
Tsuolbmajavri	68.68	22.08	P	T_{Jul}	1.21	0.86	0.25	0	1.80	Seppä and Birks, 2001
Tsuolbmajavri	68.68	22.08	D	T_{Jul}	0.27	0.81	0.26	0	1.69	Korhola et al., 2000
Tsuolbmajavri	68.68	22.08	C (bummer)	T_{Jul}	−0.20	0.40	0.29	0.05	0.99	Korhola et al., 2002
Tsuolbmajavri	68.68	22.08	C (WA_PLS)	T_{Jul}	0.40	1.00	0.19	0.05	2.04	Korhola et al., 2002
Torneträsk	68.50	19.00	T	T_{JJA}	−0.55	0.37	0.09	0	0.77	Grudd et al., 2002
Vuoskkujavri	68.33	19.10	P	T_{Jul}	1.25	1.25	0.11	0	2.51	Bigler et al., 2002
Vuoskkujavri	68.33	19.10	D	T_{Jul}	1.87	1.15	0.16	0	2.33	Bigler et al., 2002
Vuoskkujavri	68.33	19.10	C	T_{Jul}	0.42	1.61	0.37	0	3.29	Bigler et al., 2002
Vuoskkujavri	68.33	19.10	P	T_{Jan}	−2.25	3.00	0.78	0	6.21	Bigler et al., 2002
Lake 850	68.37	19.12	D	T_{Jul}	−0.12	0.98	0.11	0	1.97	Larocque and Bigler, 2004
Lake 850	68.37	19.12	C	T_{Jul}	−0.21	0.91	0.12	0	1.83	Bigler et al., 2003
Lake Njulla	68.37	18.70	D	T_{Jul}	0.73	0.99	0.35	0	2.10	Bigler et al., 2003
Lake Njulla	68.37	18.70	C	T_{Jul}	1.65	1.05	0.42	0	2.25	Bigler et al., 2003
Vuolep Njakajaure	68.33	18.78	D	T_{Jul}	1.19	1.39	0.53	0	2.97	Bigler et al., 2006
Lake Tibetanus	68.33	18.70	P	T_{Jul}	3.18	1.11	0.29	0	2.83	Hammarlund et al., 2002
Chuna Lake	67.95	32.48	P	T_{Jul}	1.30	0.82	0.06	0.02	1.64	Solovieva et al., 2005
Sjuodjijaure	67.37	18.07	P	T_{Jul}	1.15	1.70	0.13	0	2.45	Rosén et al., 2001
Sjuodjijaure	67.37	18.07	D	T_{Jul}	1.55	0.89	0.04	0	1.79	Rosén et al., 2001
Sjuodjijaure	67.37	18.07	C	T_{Jul}	1.43	1.06	0.12	0	2.14	Rosén et al., 2001
Jeknajaure	67.22	17.80	D	T_{Jul}	2.94	0.89	0.62	0	2.17	Rosén et al., 2003
Niak	67.50	18.07	D	T_{Jul}	1.00	0.74	0.17	0	1.51	Rosén et al., 2003
Seukokjaure	67.77	17.52	P	T_{Jul}	0.53	1.48	0.14	0	2.96	Rosén et al., 2003
Seukokjaure	67.77	17.52	D	T_{Jul}	−0.08	1.13	0.09	0	2.27	Rosén et al., 2003
Seukokjaure	67.77	17.52	C	T_{Jul}	−0.16	1.15	0.03	0	2.95	Rosén et al., 2003
Svanåvatnet	66.42	14.05	P	T_{Jul}	2.16	1.14	0.13	0	2.28	Bjune and Birks, 2008
Svanåvatnet	66.42	14.05	P	T_{Jan}	1.90	2.71	0.75	0	5.62	Bjune and Birks, 2008
Søylegrotta	66.62	13.68	S	T_{Ann}	0	1.80	0	0.06	3.60	Lauritzen and Lundberg, 1999
Lake Berkut	66.35	36.67	C	T_{Jul}	−0.27	1.29	0.08	0	2.59	Illashuk et al., 2005
Lake Spåime	63.12	12.32	C	T_{Jul}	1.46	1.21	0.62	0	2.71	Hammarlund et al., 2004
Råtasjøen	62.27	9.83	C	T_{Jul}	0.20	1.44	0.29	0.07	2.92	Velle et al., 2005
Brurskardstjørni	61.42	8.67	C	T_{Jul}	0.80	1.01	0.33	0.07	2.13	Velle et al., 2005
Finse stationsdamm	60.60	7.50	C	T_{Jul}	1.40	1.01	1.00	0.09	2.86	Velle et al., 2005
Holebudalen	59.83	6.98	C	T_{Jul}	0.55	1.01	0.85	0.07	2.65	Velle et al., 2005
Vestre Øykjamyrttjørn	59.82	6.00	C	T_{Jul}	−0.10	1.01	0.78	0.07	2.55	Velle et al., 2005
Vestre Øykjamyrttjørn	59.82	6.00	p	T_{Jul}	0.95	0.96	0.32	0	2.02	Bjune et al., 2005

Table 1. (a) Continued.

Site	Lat N	Lon E	Proxy[a]	Variable[b]	6 ka–0 ka	σ_c	σ_d	σ_r	$2\sigma_{tot}$	Reference
Trettetjørn	60.72	7.00	P	T_{Jul}	2.54	1.01	0.32	0	2.12	Bjune et al., 2005
Laihalampi	61.48	26.07	P	T_{Ann}	1.65	0.74	0.46	0	1.73	Heikkilä and Seppä, 2003
Arapisto	60.58	24.08	P	T_{Ann}	2.71	1.01	0.08	0	2.02	Sarmaja-Korjonen and Seppä, 2007
Gilltjärnen	60.08	15.83	C	T_{Jul}	−0.36	0.89	0.19	0	1.81	Antonsson et al., 2006
Gilltjärnen	60.08	15.83	P	T_{Ann}	0.93	0.93	0.30	0	1.95	Antonsson et al., 2006
Trehörningen	58.55	11.60	P	T_{Ann}	1.40	1.09	0.20	0	2.22	Antonsson and Seppä, 2007
Flarken	58.55	13.67	P	T_{Ann}	2.05	0.93	0.20	0	1.89	Seppä et al., 2005
Lake Raigastvere	58.58	26.65	P	T_{Ann}	2.60	0.78	0.42	0	1.77	Seppä and Poska, 2004
Lake Viitna	59.45	26.08	P	T_{Ann}	2.60	0.78	0.12	0	1.58	Seppä and Poska, 2004
Lake Ruila	59.17	24.43	P	T_{Ann}	3.50	1.08	0.05	0	2.16	Seppä and Poska, 2004
Iceland										
Torfaldsvatn	66.06	−20.38	C	T_{Jul}	0.06	1.55	0.19	0	3.13	Axford et al., 2007
St Vidarvatn	66.23	−15.84	C	T_{Jul}	0.23	1.55	0.50	0	3.27	Axford et al., 2007
Greenland										
GISP2	75.60	−38.50	ice $\delta^{18}O$	T_{Ann}	1.67	?	0	0	?	Alley et al., 2000
Dye 3	65.20	−43.80	Bt	T_{Ann}	1.82	0.30	0	0	0.59	Dahl-Jensen et al., 1998
GRIP	72.60	−37.60	Bt	T_{Ann}	2.57	0.30	0	0	0.59	Dahl-Jensen et al., 1998
N. America										
KR02	71.34	113.78	P (PLS)	T_{Jul}	−0.16	0.75	0.24	0	1.57	Peros and Gajewski, 2008
KR02	71.34	113.78	P (WAPLS)	T_{Jul}	−0.19	0.71	0.21	0	1.49	Peros and Gajewski, 2008
KR02	71.34	113.78	P (MAT)	T_{Jul}	−0.20	0.72	0.03	0	1.44	Peros and Gajewski, 2008
Dyer Lower	66.62	−61.65	P (RESP)	T_{Jul}	0.30	0.13	0.02	0.13	0.26	Kerwin et al., 2004
Dyer Lower	66.62	−61.65	P (MAT)	T_{Jul}	0.07	0.13	0.02	0.13	0.26	Kerwin et al., 2004
Zagoskin Lake	63.44	−161.90	C	T_{Jul}	0.48	1.96	0.03	0	3.93	Kurek et al., 2009
Iglutalk Lake	66.14	−66.08	P (RESP)	T_{Jul}	0.55	0.03	0.02	0.03	0.06	Kerwin et al., 2004
Iglutalk Lake	66.14	−66.08	P (MAT)	T_{Jul}	0.25	0.03	0.02	0.03	0.06	Kerwin et al., 2004
Burial Lake	68.44	−158.83	C	T_{Jul}	0.18	1.95	0.15	0	3.92	Kurek et al., 2009
Lake Vhc1	60.78	−69.83	P (RESP)	T_{Jul}	−0.20	0.71	0.16	0.06	1.43	Kerwin et al., 2004
Lake Vhc1	60.78	−69.83	P (MAT)	T_{Jul}	0.40	0.71	0.16	0.06	1.43	Kerwin et al., 2004
Lake LRI	58.58	−75.25	P	T_{Jul}	0.11	1.20	0.16	0.04	2.37	Sawada et al., 1999
U. Fly Lake	61.07	138.09	P	T_{Jul}	1.60	1.23	0.43	0	2.62	Bunbury and Gajewski, 2009
U. Fly Lake	61.07	138.09	C (MAT)	T_{Jul}	1.00	1.55	0.06	0	3.11	Bunbury and Gajewski, 2009
U. Fly Lake	61.07	138.09	C (WAPLS)	T_{Jul}	1.20	1.59	0.04	0	3.18	Bunbury and Gajewski, 2009
Lake JR01	69.90	−95.07	P	T_{Jul}	2.44	0.08	0.50	0	1.00	Zabenskie and Gajewski, 2007
Lake CF3	70.53	−68.37	C	T_{Jul}	0.48	2.24	0.01	0	4.48	Briner et al., 2007
Lake CF8	70.56	−68.95	C	T_{Jul}	0.42	1.67	0.03	0	3.35	Axford et al., 2009

Table 1. (a) Continued.

Site	Lat N	Lon E	Proxy[a]	Variable[b]	6 ka–0 ka	σ_c	σ_d	σ_r	$2\sigma_{tot}$	Reference
Siberia										
Lyadhej-To	68.25	65.75	C	T_{Jul}	1.00	1.13	0.20	0	2.29	Andreev et al., 2005
Lyadhej-To	68.25	65.75	P	T_{Jul}	2.18	2.12	0.33	0.09	4.30	Andreev et al., 2005
Khaipudurskaya	68.00	60.00	P	T_{Jul}	3.11	0.85	1.28	0.08	3.01	Andreev and Klimanov, 2000
Khaipudurskaya	68.00	60.00	P	T_{Jan}	1.94	1.41	1.06	0.08	3.53	Andreev and Klimanov, 2000
Khaipudurskaya	68.00	60.00	P	T_{Ann}	2.9	0.85	1.17	0.08	2.90	Andreev and Klimanov, 2000
Taymyr	70.77	99.13	P	T_{Jul}	3.99	0.85	0.23	0.08	1.73	Andreev and Klimanov, 2000
Taymyr	70.77	99.13	P	T_{Jan}	2.46	1.41	0.18	0.08	2.80	Andreev and Klimanov, 2000
Taymyr	70.77	99.13	P	T_{Ann}	2.34	0.85	0.12	0.08	1.69	Andreev and Klimanov, 2000
Kazaché	70.77	136.25	P	T_{Jul}	0.44	0.85	1.01	0.12	2.65	Andreev et al., 2001
Kazaché	70.77	136.25	P	T_{Jan}	3.23	1.41	0.46	0.12	2.98	Andreev et al., 2001
Kazaché	70.77	136.25	P	T_{Ann}	3.00	0.85	0.39	0.12	1.88	Andreev et al., 2001
Levison-Lessing	74.47	98.63	P	T_{Jul}	2.25	0.60	0.35	0.07	1.40	Andreev et al., 2003
Levison-Lessing	74.47	98.63	P	T_{Jan}	3.00	1.00	0.50	0.07	2.24	Andreev et al., 2003
Levison-Lessing	74.47	98.63	P	T_{Ann}	1.32	0.60	0.40	0.07	1.45	Andreev et al., 2003
Lama Lake	69.53	90.20	P (PFT)	T_{Jul}	2.28	0.85	0.32	0.06	1.82	Andreev et al., 2004
Lama Lake	69.53	90.20	P (IS)	T_{Jul}	3.00	0.85	0.18	0.06	1.74	Andreev et al., 2004
Lama Lake	69.53	90.20	D	T_{Jul}	0.98	0.92	0.31	0	1.95	Kumke et al., 2004
Marine										
CR 948/2011	66.97	7.64	D	SST_{Aug}	2.64	0.80	0.65	2.05	2.88	Birks and Koç 2002, Andersen et al., 2004
JR51-GC35	67.00	17.96	A	SST_{Ann}	0.89	1.14	1.55	0	3.85	Bendle and Rosell-Melé, 2007
MD95-2011, JM97-948/2A BC	66.97	7.64	F	SST_{Sum}	−0.31	1.04	0.14	0	2.09	Andersson et al., 2003
T88-2, JM01-1199	71.99	14.36	F	SST_{Sum}	−1.08					Hald et al., 2007
MD99-2269	66.85	−20.85	D	SST_{Aug}	−0.21	0.74	0.50	0	1.77	Justwan et al., 2008
CR 19/05	67.13	−30.90	F	SST_{Aug}	0.75	0.92	0.45	0.09	2.06	Andersen et al., 2003
B997-324	66.89	−18.98	F	SST_{Jul}	−0.52	0.79	0.11	0	1.60	Smith et al., 2005
B997-321	66.53	−21.50	F	SST_{Jul}	0.91	0.89	0.35	0	1.92	Smith et al., 2005
91-039	77.27	−74.33	Dc	SST_{Aug}	−1.24					Levac et al., 2001
B997-347	63.93	−24.48	F	SST_{Jul}	−1.62	0.88	0.10	0	1.78	Smith et al., 2005
MD99-2304	77.62	9.95	F	SST_{Sum}	−1.24	1.6	0.25	0.18	3.26	Hald et al., 2007

Table 1. (b) Precipitation reconstructions from proxy data north of 58° N discussed in this paper.

Site	Lat N	Lon E	Proxy[a]	Variable[b]	6 ka–0 ka	σ_c	σ_d	σ_r	$2\sigma_{tot}$	Reference
Fennoscandia										
Dalmutladdo	69.17	20.72	P	P_{Ann}	−188	217	18	0	541	Bjune et al., 2004
Aspvatnet	69.73	19.98	Sd	P_{win}	−7	?	?	?	?	Bakke et al., 2005
Toskaljavri	69.20	21.47	P	P_{Ann}	276	271	189	0	661	Seppä and Birks, 2002
Tsuolbmajavri	68.68	22.08	P	P_{Ann}	−231	296	139	0	654	Seppä and Birks, 2001
Vuoskkujavri	68.33	19.10	P	P_{Ann}	−168	472	57	0	952	Bigler et al., 2002
Lake Tibetanus	68.33	18.70	P	P_{Ann}	244	443	128	0	922	Hammarlund et al., 2002
Chuna Lake	67.95	32.48	P	P_{Ann}	−107	48	5	0	96	Solovieva et al., 2005
Svanåvatnet	66.42	14.05	P	P_{Ann}	326	383	157	0	828	Bjune and Birks, 2008
Søylegrotta	66.62	13.68	S	T_{Ann}	0	1.80	0	0.06	3.60	Lauritzen and Lundberg, 1999
Vestre Øykjamyrttjørn	59.82	6.00	P/ELA	P_{win}	48	?			?	Bjune et al., 2005
Trettetjørn	60.72	7.00	P/ELA	P_{win}	−24	?			?	Bjune et al., 2005
Hardangerjøkulen	60.5	7.71	P/ELA	P_{win}	−56	?			?	Bjune et al., 2005
Jostedalsbreen	61.58	7.50	P/ELA	P_{win}	−56	?			?	Bjune et al., 2005
N. Folgefonna	60.23	6.42	P/ELA	P_{win}	−31	?			?	Bjune et al., 2005
N. America										
KR02	71.34	−113.78	P (PLS)	P_{Ann}	3	24	1	0	47	Peros and Gajewski, 2008
KR02	71.34	−113.78	P (WAPLS)	P_{Ann}	6	22	2	0	45	Peros and Gajewski, 2008
KR02	71.34	−113.78	P (MAT)	P_{Ann}	−1	20	0	0	39	Peros and Gajewski, 2008
U. Fly Lake	61.07	−138.09	P	P_{Ann}	−55	202	8	0	397	Bunbury and Gajewski, 2009
Siberia										
Lyadhej-To	68.25	65.75	P	P_{Ann}	118	86	20	6	177	Andreev et al., 2005
Khaipudurskava	68.00	60.00	P	P_{Ann}	77.8	35	22	4	84	Andreev and Klimanov, 2000
Taymyr	70.77	99.13	P	P_{Ann}	143	35	4	4	70	Andreev and Klimanov, 2000
Kazaché	70.77	136.25	P	P_{Ann}	13	35	21	7	83	Andreev et al., 2001
Levison-Lessing	74.47	98.63	P	P_{Ann}	74	25	0	2	50	Andreev et al., 2003
Lama Lake	69.53	90.20	P (PFT)	P_{Ann}	−3	35	4	2	71	Andreev et al., 2004
Lama Lake	69.53	90.20	P (IS)	P_{Ann}	51	35	21	2	82	Andreev et al., 2004

[a] P, pollen; D, diatoms; Ch, chironomids; T, tree-ring with ELA; S, speleothem δ^{18}O; I, ice δ^{18}O; Bt, Borehole temp; Sd, sediment density; F, foraminifera; Dc, dinocysts

[b] T_{Ann}, Annual mean temp (°C); T_{Jul}, July mean temp (°C); T_{Jan} January mean temp (°C)

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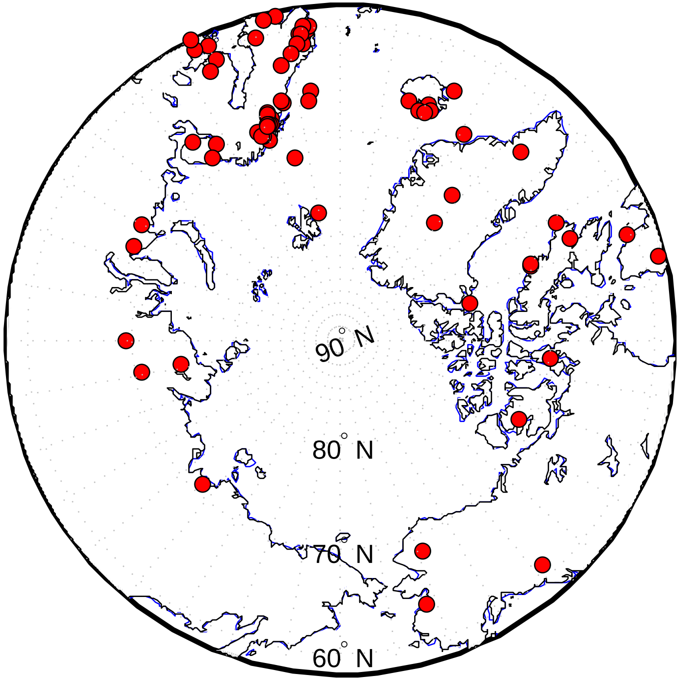

Fig. 1. Map of the sites from which quantitative reconstructions of proxy data have been collected.

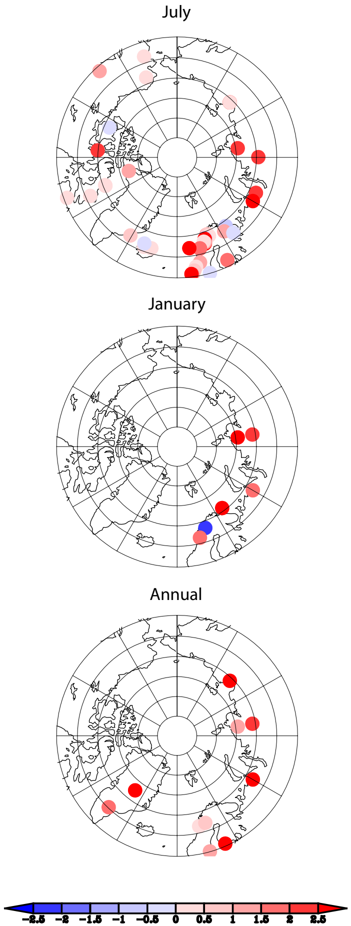

Fig. 2. Overview of the average change in temperature (annual, summer, winter) between 6 ka and 0 ka at the different proxy sites over the northern high latitudes.

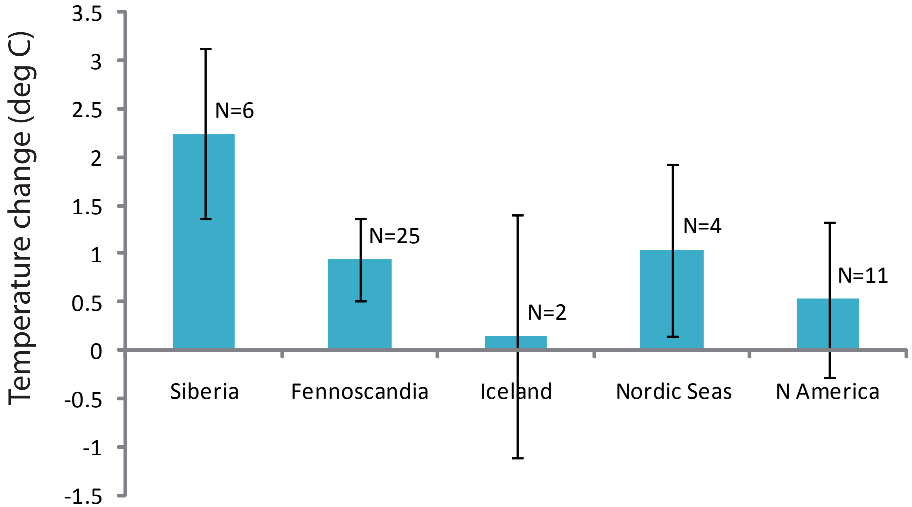

Fig. 3. Average change in summer temperature (6 ka–0 ka) with error bars (2σ) for different regions. N=number of sites that have contributed to the average.

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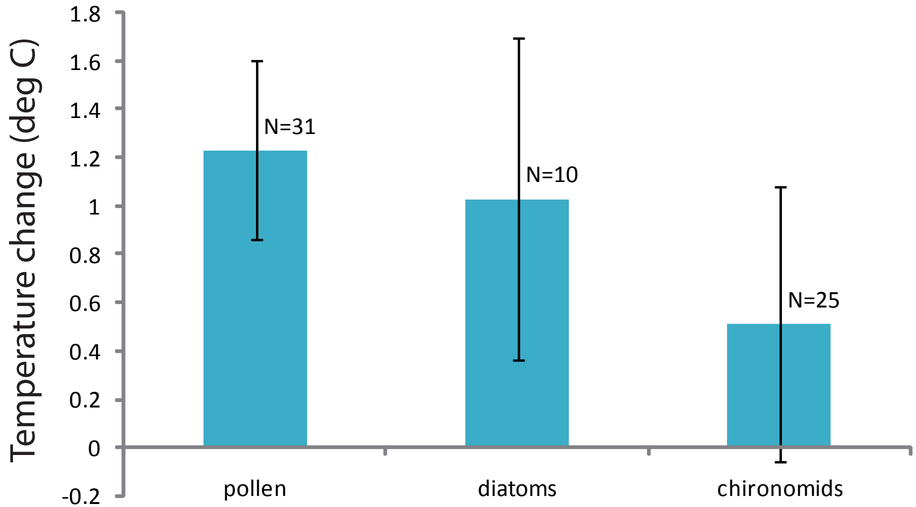

Fig. 4. Average change in July temperature (6 ka–0 ka) with error bars (2 sigma) in pollen, diatoms and chironomids. *N*=number of reconstructions that have contributed to the average.

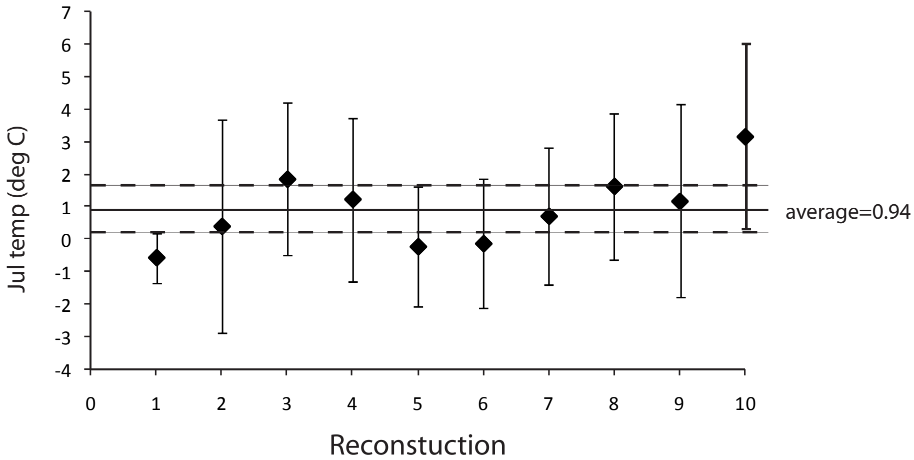

Fig. 5. Change in summer temperature (6 ka–0 ka) from 10 reconstructions in the Abisko area, northern Sweden. Error bars (2σ) for each reconstruction and for the total average (dotted lines) are indicated. 1) Torneträsk, tree-ring width, 2) Vuoskkujavri, chironomids, 3) Vuoskkujavri, diatoms, 4) Vuoskkujavri, pollen, 5) Lake 850, chironomids, 6) Lake 850, diatoms, 7) Lake Njulla, diatoms, 8) Lake Njulla, chironomids, 9) Voulep Njakajaure, diatoms, 10) Lake Tibetanus, pollen. Note that all the reconstructions are reconstructions of July temperature except the tree-ring reconstruction which is a reconstruction of July–August temperature.