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A regional approach to the medieval warm period and the little ice age

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1. Introduction

In order to gain knowledge of the temperature variability prior to the establishment of a widespread network of instrumental measurements c. AD 1850, we have to draw information from proxy data sensitive to temperature variations. Such data can be extracted from various natural recorders of climate variability, such as corals, fossil pollen, ice-cores, lake and marine sediments, speleothems, and tree-ring width and density, as well as from historical records (for a review, see IPCC 2007; Jones et al. 2009; NRC 2006). Considerable effort has been made during the last decade to reconstruct global or northern hemispheric temperatures for the past 1000 to 2000 years in order to place the observed 20th century warming in a long-term perspective (e.g., Briffa, 2000; Cook et al., 2004; Crowley and Lowery, 2000; D'Arrigo, 2006; Esper et al., 2002; Hegerl et al., 2007; Jones et al., 1998; Jones and Mann, 2004; Juckes et al., 2007; Ljungqvist, 2010; Loehle, 2007; Mann et al., 1999; Mann et al., 2008; Mann et al., 2009; Mann and Jones, 2003; Moberg et al., 2005; Osborn and Briffa, 2006).

Less effort has been put into investigating the key question of to what extent earlier warm periods have been as homogeneous in timing and amplitude in different geographical regions as the present warming. It has been suggested that late-Holocene long-term temperature variations, such as the Medieval Warm Period (MWP) and the Little Ice Age (LIA), have been restricted to the circum-North Atlantic region (including Europe) and have not occurred synchronic in time with warm and cold periods respectively in other regions (Hughes and Diaz, 1994; Mann et al., 1999; Mann and Jones, 2003). This view has, however, been increasingly challenged through the ever growing amount of evidence of a global (or at least northern hemispheric) extent of the MWP and the LIA that have become available (see, for example, Esper and Frank, 2009; Ljungqvist, 2009, 2010; Moberg et al., 2005; Wanner et al., 2008).

A main obstacle in large-scale temperature reconstructions continues to be the limited and unevenly distributed number of quantitative palaeotemperature records extending back a millennium or more. The limited number of records have rendered it impossible to be very selective in the choice of data. Palaeotemperature records used in a large-scale temperature reconstruction should preferably be accurately dated, have a high sample resolution and have a high correlation with the local instrumental temperature record in the calibration period (see the discussion in Jones et al., 2009). The number of long quantitative

palaeotemperature records from across the globe, of which a majority are well suited for being used in large-scale temperature reconstructions, have been rapidly increasing in recent years (Ljungqvist, 2009). Thus, it has now become possible to make regional temperature reconstructions for many regions that can help us to assess the spatio-temporal pattern and the MWP and LIA. Only by a regional approach can we truly gain an understanding of the temperature variability in the past 1–2 millennia and assess the possible occurrence of globally coherent warm and cold periods. Presently, only four regional multi-proxy temperature reconstructions exist: two for eastern Asia (Yang et al., 2002; Ge et al., 2010), one for the Arctic (Kaufman et al., 2009), and one for South America (Neukom et al., 2010). Six new quantitative regional multi-proxy temperature reconstructions will here be presented in order to improve our understanding of the regional patterns of past temperature variability.
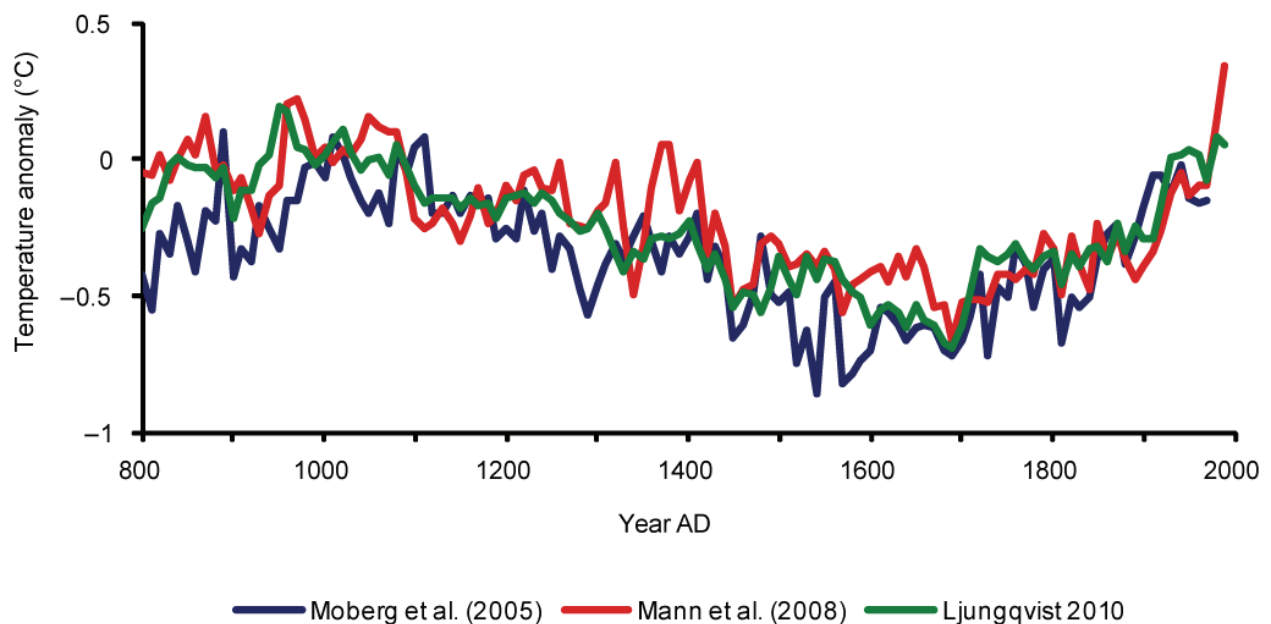

Fig. 1. Comparison of three recent millenium-long multi-proxy Northern Hemisphere temperature reconstructions: decadal means of Moberg et al. (2005), the 'error-in-variables' (EIV) regression method variant of Mann et al. (2008), and the extra-tropical Northern Hemisphere reconstruction by Ljungqvist (2010).

2. New regional temperature reconstructions

Only for limited parts of the Northern Hemisphere is the data coverage sufficient for making quantitative regional temperature reconstructions in order to assess the regional pattern of temperature changes during the last 12 centuries. This has been done for: 1) warm season temperatures of Scandinavia north of 60°N, 2) warm season temperatures for northern Siberia, 3) annual mean temperatures for Greenland, 4) warm season temperatures for the Alp region of Central Europe, 5) annual mean temperatures for China, and 6) annual mean temperatures for the whole of the North American continent. The reconstructions follow the simple but robust method outlined in the new multi-proxy temperature reconstruction for the extra-tropical (90°–30°N) Northern Hemisphere by Ljungqvist (2010). Only proxy records with reasonably high resolution (multi-decadal or better) were utilised

and records with lower resolution were instead used for the purpose of verifying the reconstructed low-frequency trends of the reconstructions.

We use the common "composite-plus-scale" method (Lee et al. 2008; von Storch et al. 2004) for creating the different regional temperature reconstructions. First, all records with less than annual resolution were linearly interpolated to have annual resolution and then calculated 10-year-mean values of each record. All the 10-year-mean values were normalized to zero mean and unit standard deviation fitting the decadal mean and variance AD 1000–1899. The arithmetic mean of all the normalized records included in each regional reconstruction was then calculated. Each regional reconstruction was scaled to fit the decadal mean and variance over the longest possible time period in the variance adjusted CRUTEM3 instrumental temperature record (Brohan et al., 2006) and adjusted to have a zero equaling the 1961–1990 mean of this instrumental record. A 2 standard deviation error bar for each regional reconstruction was calculated from the decadal correlation between proxy and instrumental temperature in the calibration period.

Proxy location	Latitude	Longitude	Proxy type*	Sample resolution	Season bias	Reference
1. Finnish Lapland	69.00	25.00	TRW	Annual	Summer	Helama et al. 2009
2. Lake Tsuolbmajavri	68.45	22.05	Lf	Multi-decadal	Summer	Korhola et al. 2000
3. Torneträsk	68.31	19.80	MXD	Annual	Summer	Grudd 2008
4. Jämtland	63.10	13.30	TRW	Annual	Summer	Linderholm and Gunnarson 2005
5. Severnaja	81.00	106.00	Lf	Decadal	Summer	Solomina and Alverson 2004
6. Taimyr peninsula	73.00	105.00	TRW	Annual	Summer	Naurzbaev et al. 2002
7. Indigirka	70.00	149.00	TRW	Annual	Summer	Moberg et al. 2005
8. Yamal			TRW	Annual	Summer	Briffa 2000
9. Polar Urals	66.83	65.75	MXD	Annual	Summer	Esper et al. 2002a
10. Lower Murray Lake	81.21	–69.32	V	Annual	Summer	Cook et al. 2009
11. GISP	72.60	–38.50	Is	Annual	Annual	Johnsen et al. 2001
12. GRIP	72.35	–37.38	Is	Annual	Annual	Grootes and Stuiver 1997
13. Crête	71.12	–37.32	Is	Annual	Annual	Clausen, et al. 1988
14. Nansen Fjord	68.25	–29.60	Sd	Decadal	Summer	Jennings and Weiner 1996
15. Igaliku Fjord	60.40	–46.00	Sd	Decadal	Summer	Jensen et al. 2004
16. Lake Silvaplana	46.45	9.48	Lf	Annual to decadal	Summer	Larocque et al. 2009
17. The Alps	46.30	8.00	MXD	Annual	Summer	Büntgen et al. 2006
18. Mongolia	48.30	98.93	TRW	Annual	Summer	D'Arrigo et al. 2001
19. Shihua Cave	39.47	115.56	Sp	Annual	Summer	Tan et al. 2003
20. Dulan	36.00	98.00	TRW	Annual	Annual	Zhang et al. 2003

21. E. China	35.00	114.00	D	Decadal	Annual	Yang et al. 2002
22. E. China	35.00	114.00	D	Decadal	Winter	Ge et al. 2003
23. Wanxiang	33.19	105.00	Sp	Decadal	Annual	Zhang et al. 2008
24. Zhang Delta	32.00	121.00	D	Decadal	Annual	Zhang et al. 2008
25. Hesheng	19.41	109.36	Sp	Decadal	Annual	Hu et al. 2008
26. Devon Island	75.33	−82.50	Ice-core	5 years	Annual	Fisher et al. 1983
27. Iceberg Lake, Alaska	60.78	−142.95	Lf	Annual	Summer	Loso 2009
28. Gulf of Alaska	60.00	−145.00	MXD	Annual	Summer	D'Arrigo et al. 2006
29. Canadian Rockies	52.15	−117.15	MXD	Annual	Summer	Luckman and Wilson 2005
30. Chesapeake Bay	39.00	−76.40	Md	Multi-decadal	Spring	Cronin et al. 2003
31. Bermuda Rise	33.72	−57.63	Md	Multi-decadal	Annual	Keigwin 1996
32. Nicoa Cave	10.20	−85.30	Sp	Decadal	Annual	Mann et al. 2008
33. Punta Laguna	20.63	−87.50	Lf	Decadal	Annual	Curtis et al. 1996

* D, documentary; Lf, lake/river fossils and sediments; MXD, tree-ring maximum latewood density; Md, marine sediments; Sp, speleothem isotopic analysis; TRW, tree-ring width; V, varved thickness sediments.

Table 1. All temperature proxy records used in the regional temperature reconstructions with information regarding geographical location, latitude and longitude, type of proxy, sample resolution, season bias, and reference to the original publication. The geographical locations of the records are shown on the map in Fig. 2.

Proxy location	Latitude	Longitude	Proxy type*	Sample resolution	Season bias	Reference
A. Lake Gammelheimenvatnet	68.47	17.75	Po	Centennial	Summer	Bjune et al. 2009
B. Lake Sjuodjijaure	67.37	18.07	Lf	Centennial	Summer	Rosén et al. 2003
C. Korallgrottan	64.89	14.16	Sp	Multi-decadal	Annual	Sundqvist et al. 2010
D. Taimyr pollen	70.77	99.13	Po	Centennial	Summer	Andreev et al. 2002
E. Indigirka pollen	70.00	149.00	Po	Centennial	Summer	Velichko et al. 1997
F. Yamal tree-line	67.00	69.00	O	Centennial	Summer	Solomina and Alverson 2004
G. GRIP	72.35	−37.38	B	Centennial	Annual	Dahl-Jensen et al. 1998
H. Dye-3	65.11	−43.49	B	Centennial	Annual	Dahl-Jensen et al. 1998
I. Qipisarqo Lake	61.01	−47.75	Lf	Centennial	Summer	Kaplan et al. 2002
J. Lake Neuchatel	46.80	6.70	Po	Centennial	Summer	Filippi et al. 1999
K. Aletsch Glacier	46.38	7.75	O	Centennial	Summer	Holzhauser et

						al. 2005
L. Groner Glacier	46.05	7.62	O	Centennial	Summer	Holzhauser et al. 2005
M. Lake Qinghai	37.00	100.00	Lf	Multi-decadal	Annual	Liu et al. 2006
N. Hongyuan	32.46	102.3	Lf	Centennial	Annual	Yang et al. 2002
O. Jiaming Lake	25.01	121.3	Lf	Centennial	Annual	Yang et al. 2002
P. Farewell Lake	62.55	–153.63	Lf	Centennial	Summer	Hu et al. 2001
Q. Tebenkof Glacier	60.75	–148.45	O	Centennial	Summer	Barclay et al. 2009
R. North America pollen stack	70–30	–50–170	Po	Centennial	Summer	Viau et al. 2006

* B, borehole; Lf, lake/river fossils and sediments; Md, marine sediments; Sp, speleothem isotopic analysis; O, other types of proxies; Po, pollen.

Table 2. All temperature proxy records used for verifying the low-frequency trends in the regional temperature reconstruction with information regarding geographical location, latitude and longitude, type of proxy, sample resolution, season bias, and reference to the original publication. The geographical locations of the records are shown on the map in Fig. 2.

2.1 Scandinavia

Scandinavia is probably the most data rich region in the world when it comes to palaeotemperature proxy data. Climate and environmental research has a long history in Scandinavia and numerous studies of past climate and vegetation have been conducted, especially in the far north of Scandinavia, for several decades. Many different kinds of archives have been utilised including, but not limited to, tree-ring width and density data, pollen profiles, chironomid records, diatom records, annually-laminated sediments, radiocarbon-dated megafossil tree-remains, and speleothem $\delta^{18}O$ records. Most of this data are, unfortunately, not available from databases and the majority of the records do not, moreover, possess such a high resolution that they are suited for being used in calibrated temperature reconstructions of the last 1200 years. Almost all palaeotemperature proxy data from Scandinavia, especially northern Scandinavia, primarily capture growing season temperatures and are hence biased towards the warm part of the year. In southern Scandinavia the growing season lasts approximately five months (May to September), whereas the growing season in northern Scandinavia lasts three months (June to August) or less on high elevations. We still have a limited possibility to reconstruct cold season or annual mean temperatures for Scandinavia despite the fact that the region is so comparatively rich in data.

Presently, only four records possess such quality that we can use them here for reconstructing the warm season temperature variability in Scandinavia for the last 1200 years: 1) the Torneträsk tree-ring maximum latewood density record (Grudd, 2008), 2) the Finnish Lapland tree-ring width record (Helama et al., 2009), and 3) the Jämtland tree-ring width record (Linderholm and Gunnarson, 2005). The sediment records from Lake Tsuolbmajavri in northernmost Finland (Korhola et al., 2000) also possess such a relatively high resolution and dating control that they can be useful to include in a Scandinavian

warm season temperature reconstruction. Other records usually have too low a temporal resolution.
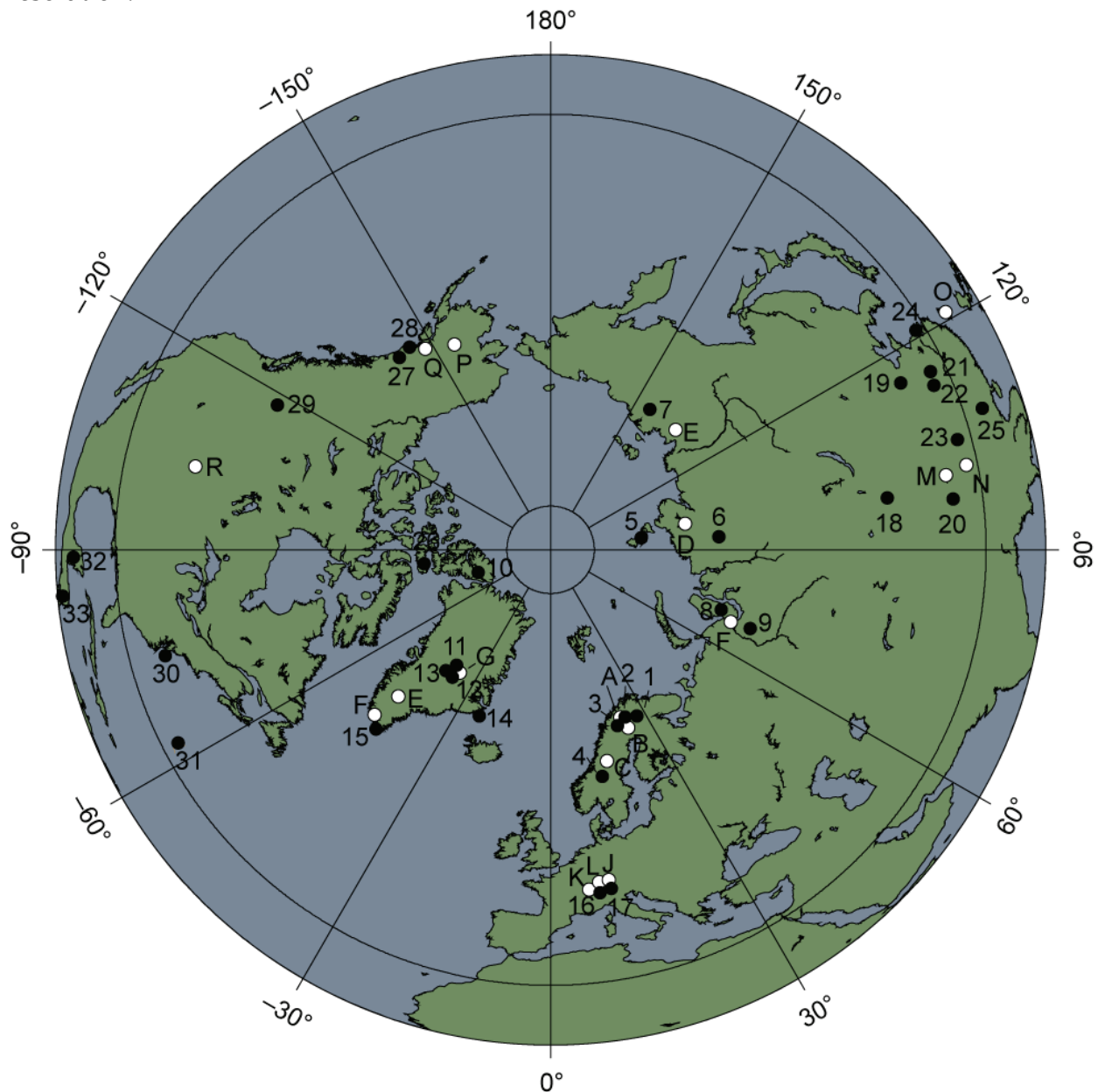

Fig. 2. Map with the geographical location of the proxy records listed in Table 1 and 2.

The Scandinavian warm season temperature reconstruction, consisting of the above-mentioned four records, was calibrated against the May to September mean temperature from the 60–70°N × 5–30°E CRUTEM3 grid cells (Fig. 3). This area represents the central and northern parts of Scandinavia. We have a long and good network of instrumental temperature measurements for Scandinavia and we can thus calibrate the reconstruction over the whole period AD 1850–1999. The correlation coefficient over the calibration period amounts to 0.88 (r^2 = 0.77). Late 20th century warm season temperatures in Scandinavia do not seem to be remarkably warm in a long-term perspective. During the MWP, occurring here c. AD 900–1100, Scandinavian warm season temperatures seem to have exceeded those

of recent decades. Temperatures also equalled or exceeded the 1961–1990 reference level in the early 15th century. A LIA cooling is clearly seen approximately AD 1560–1720 and low temperatures are also prevailing c. AD 1350 and c. AD 1900. The total reconstructed decadal temperature variability of the last 12 decades is about 2.5°C, with a centennial variability of as much as 1.5°C.
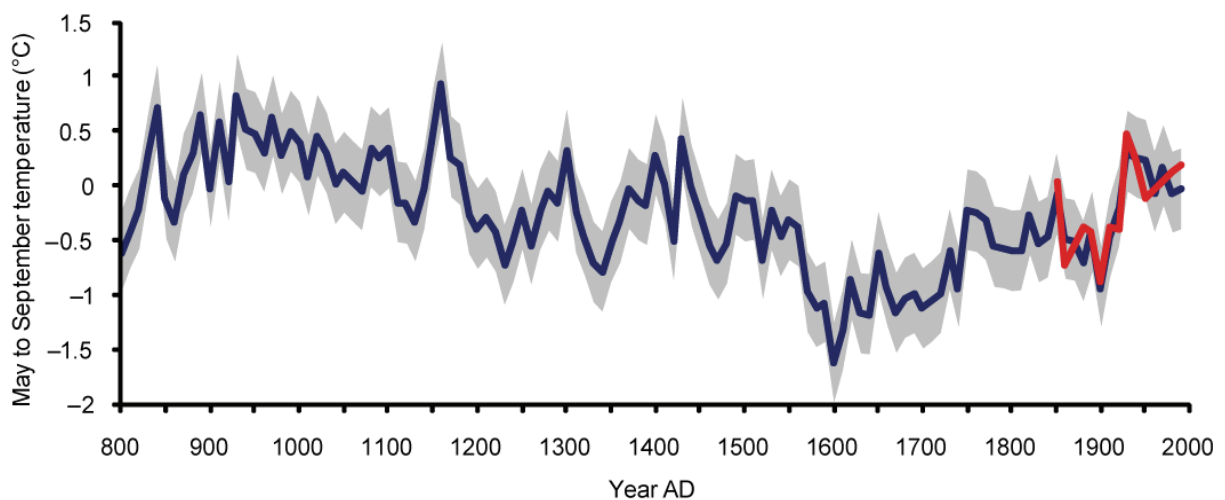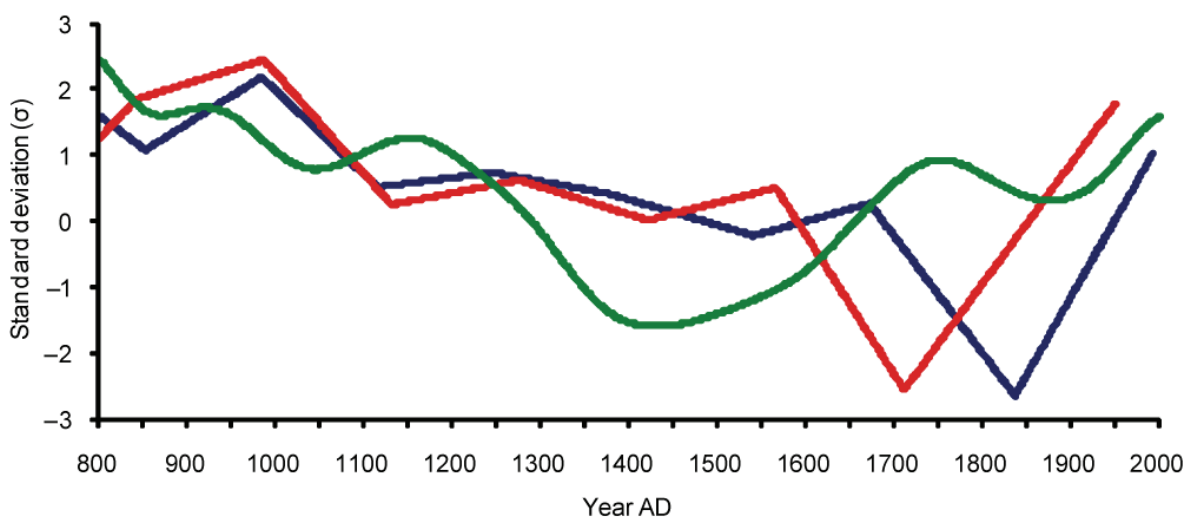

Fig. 3. May to September temperature reconstruction for Scandinavia north of 60°N (blue line) calibrated to instrumental temperatures for the same region (red line) with 2 standard deviation error bars (grey shading).

Fig. 4. Three normalised temperature records from Scandinavia shown relative to their AD 1000–1899 mean. The Korallgrottan record has been smoothed with a 100-year cubic spline filter.

As discussed above, there exist large numbers of quantitative temperature reconstructions with lower temporal resolution that can be used to compare and verify the reconstructed low-frequency trend. We have used the Lake Gammelheimenvatnet pollen-based reconstruction (Bjune et al., 2009) and the Lake Sjuodjijaure chironomid-based

reconstruction (Rosén et al., 2003) from the northern tree-line area for this purpose together with the speleothem δ18O record from Korallgrottan in Jämtland in Central Scandinavia (Sundqvist et al., 2010) (Fig. 4). Lake Gammelheimenvatnet and Lake Sjuodjijaure show high summer temperature variability whereas Korallgrottan probably reflects annual mean temperature variability. The general multi-centennial trends of the reconstruction are also seen in the three low-resolution records, and although they do not agree when the maximum LIA cooling occurred, they are consistent with regard to the occurrence of a clear MWP in Scandinavia.

2.2 Northern Siberia

Russia has a long tradition of palaeoclimatology and although much of it is still only available in the Russian language, considerable efforts were made already in the first half of the 1980s in order to incorporate Russian (then Soviet) palaeoclimatology with the Western research community (Velichko, 1984). However, the Russian palaeoclimatology has primarily been focused on long Pleistocene and Holocene perspectives and less focused on the climate variability of the last one or two millennia (Velichko et al., 1997). Siberian, as well as European Russian, pollen-based temperature reconstructions clearly show the occurrence of a MWP and a LIA but they have such a crude resolution that only multi-centennial variations can be detected (Andreev et al., 2000, 2001, 2003, 2004, 2005).

Five records, all primarily reflecting warm season temperatures, with temporal resolution high enough to be used here were found: 1) the Severnaja lake sediment record (Solomina and Alverson, 2004), 2) the Taimyr tree-ring width record (Naurzbaev et al., 2002), 3) the Indigirka tree-ring width record (Moberg et al., 2006), 4) the Yamal tree-ring width record (Briffa, 2000), and 5) the Polar Urals tree-ring maximum latewood density record (Esper et al., 2002). Regular instrumental temperature measurements were started relatively late in Siberia. Our calibration period is therefore limited to AD 1890–1989 and thus we have a limited degree of freedom. The northern Siberia warm season temperature reconstruction has been calibrated against the May to September mean temperature from the 60–80°N × 60–180°E CRUTEM3 grid cells (Fig. 5). During this 10 decade long calibration period, the correlation coefficient amounts to 0.70 (r^2 = 0.48). The relationship between the proxy composite and instrumental temperatures is thus relatively weak although the general trends are in quite good agreement.

The reconstructed Siberian warm season temperatures show that temperatures exceeded the 1961–1990 reference level c. AD 950–1150 and mostly equalled that level from c. AD 800–950 and c. AD 1150–1540. A quite clear LIA is seen c. AD 1540–1920 with temperatures about 0.5°C below the 1961–1990 reference level. Three especially distinct cold spells are seen during the LIA: c. AD 1280, a long cold period c. AD 1600–1750, and c. 1820. The overall amplitude of the reconstructed decadal variability the last 12 centuries well exceeds 1°C.

The reconstructed low-frequency temperature trend agrees well with the normalised values of three warm season temperature reconstructions with lower temporal resolution: the Yamal tree-line record (Solomina and Alverson, 2004), the Indigirka pollen-based temperature reconstruction (Velichko et al., 1997), and the Taimyr pollen-based temperature reconstruction (Andreev et al., 2002) (Fig. 6). The MWP, the LIA, and the modern warming are clearly visible in the low-resolution records, although they are less clear in the Yamal tree-line record. However, maximum LIA cooling seems to occur somewhat earlier in the

three low-resolution records than in the quantitative, calibrated temperature reconstruction (Fig. 5).
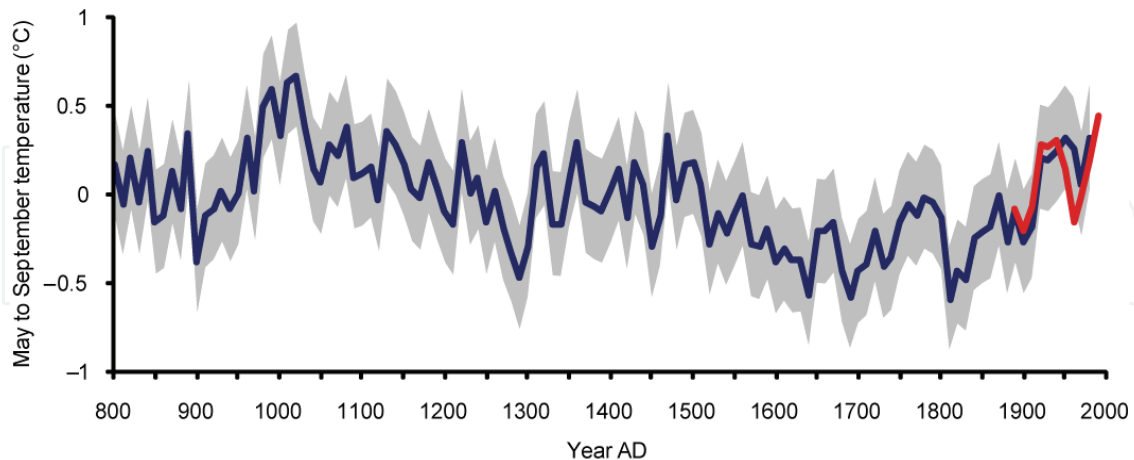

Fig. 5. May to September temperature reconstruction for northern Siberia (blue line) calibrated to instrumental temperatures for the same region (red line) with 2 standard deviation error bars (grey shading).
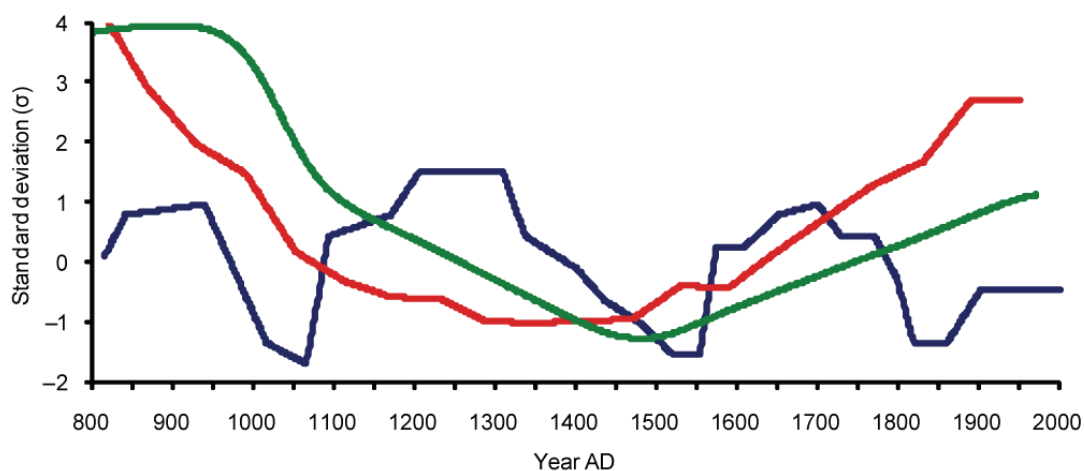

Fig. 6. Three normalised temperature records from northern Siberia shown relative to their AD 1000–1899 mean.

2.3 Greenland

Greenland has been the subject of palaeoclimatological research efforts for several decades, resulting among other things in a number of well-known $\delta^{18}O$ ice-core records (Andersen et al., 2006; Dansgaard et al., 1975). In the last decades, marine sediment cores from Greenland's extensive coastlines as well as lake sediment cores from the ice-free coastal areas have also become available (Andresen et al., 2004; Cremer et al., 2001; Moros et al. 2006; Møller et al., 2006; Roncaglia and Kuijpers, 2004; Seidenkrantz et al., 2007; Wagner and Melles, 2001). Most of the sediment records unfortunately have too low a temporal resolution and/or too large uncertainties in the dating to be useful in a quantitative multi-proxy temperature reconstruction aimed at being calibrated to temperature values. Other

potentially useful records end too early to be calibrated to instrumental temperatures (e.g. Alley, 2000). Further records are unsuitable to use in order to reconstruct Greenland's temperature since they either do not have a significant correlation to temperature (as the NorthGRIP δ^{18}O ice-core record) or stronger correlation to Icelandic rather than Greenlandic temperatures (as the Renland δ^{18}O ice-core record) (Vinther et al., 2010). Thus, all in all, we find six records useful for our purpose: 1) the Crête δ^{18}O ice-core record, 2) the GISP2 δ^{18}O ice-core record, 3) the GRIP δ^{18}O ice-core record, 4) the Nansen Fjord benthic foraminifera sea floor sediment record (Jennings and Weiner, 1996), 5) the Igaliku Fjord biostratigraphic diatom sea floor sediment record (Jensen et al., 2004), and 6) the annually varved Lower Murray Lake sediment record (Cook et al., 2009). The last record, from Lower Murray Lake, actually comes from Ellesmere Island in the northernmost Canadian Arctic Archipelago but can be used to represent climate conditions in nearby northern Greenland.

The Greenland temperature reconstruction has been calibrated against the annual mean temperature from the 85–60°N × 15–70°W CRUTEM3 grid cells (Fig. 7). The correlation coefficient over the calibration period 1850–1969 amounts to 0.74 (r^2 = 0.55). The reconstruction shows a peak value in the 10th century c. 3°C above the 1961–1990 reference level and c. 1.5°C above the mid-20th century maximum. Temperatures then gradually declined but essentially remained above the 1961–1990 reference level until about AD 1300. We can thus conclude with reasonable safety that the MWP in Greenland well exceeded the observed 20th century warming, although this did not necessarily apply to the Arctic region as a whole (Kaufman et al., 2009). Such a strong regional warming of Greenland is actually well in agreement with global temperature field reconstructions indicating a post-1990 warming exceeding the medieval one on a global scale but not on Greenland (Mann et al., 2009). A pronounced cold period occurred in the mid to late 14th century, probably marking the onset of the LIA, although temperatures then rose again and hovered around the 1961–1990 reference level until the main phase of the LIA commenced in the late 17th century and lasted until c. AD 1920. It may be noted that no late 20th century warming is visible in either the reconstructed or the instrumental temperatures since the recent warming in Greenland only started in the late 1990s and still has not exceeded the level of the mid-20th century Greenland warming (Chylek et al., 2006).
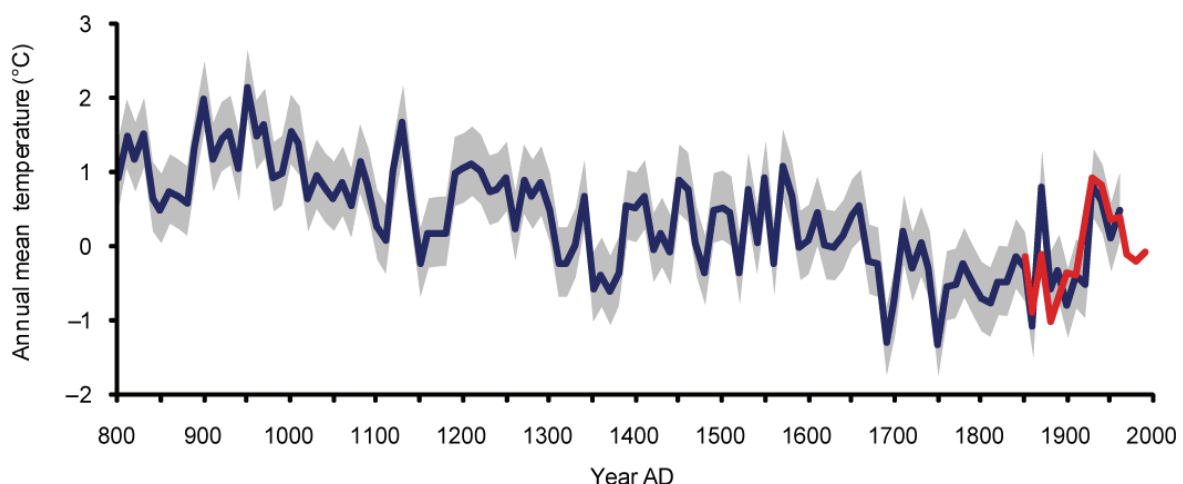

Fig. 7. Annual mean temperature reconstruction for Greenland (blue line) calibrated to instrumental temperatures for the west coast of Greenland (red line) with 2 standard deviation error bars (grey shading).
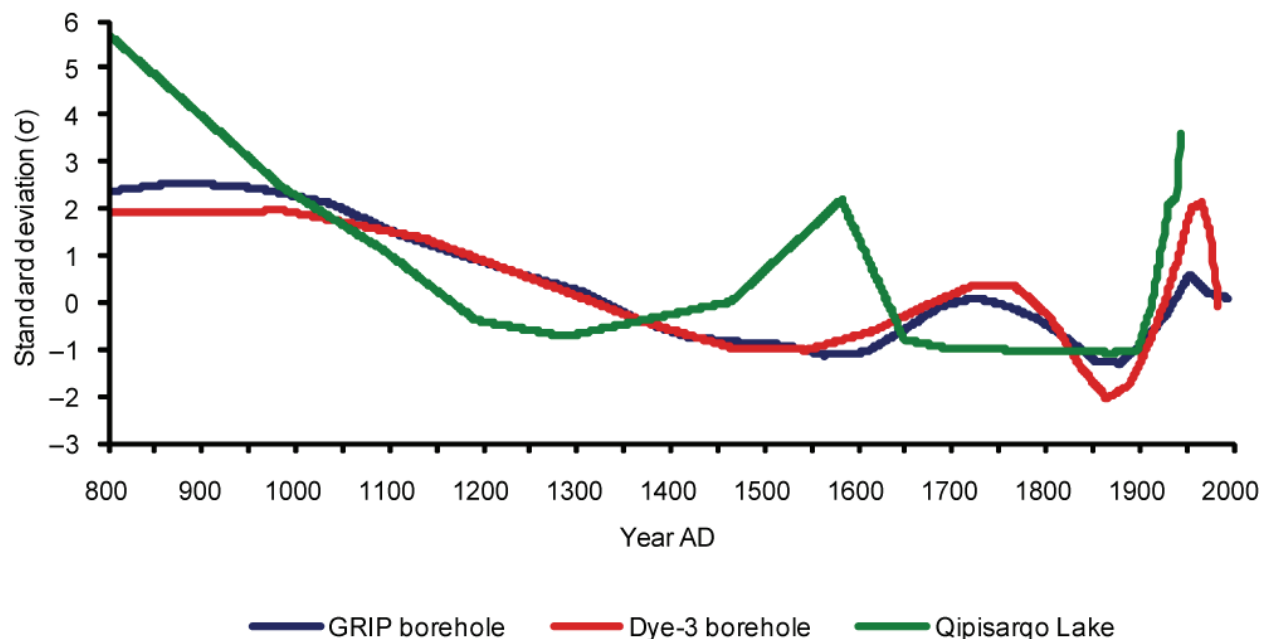
Fig. 8. Three normalised temperature records from Greenland shown relative to their AD 1000–1899 mean.

The reconstructed low-frequency temperature trend agrees well with the normalised values of the Dye-3 and GRIP deep borehole temperature profiles from the Greenland inland ice sheet (Dahl-Jensen et al., 1998) as well as with the normalised values of the biogenic silica sediment record from Qipisarqo Lake (Kaplan et al., 2002) (Fig. 8). Moreover, based on the analysis of the oxygen isotope composition of human tooth enamel ($\delta^{18}O_p$) in teeth from Norse and Inuit skeletons, Fricke et al. (1995) found a similar amplitude in the annual mean temperature variability as we have reconstructed here.

2.4 Central Europe

Considerable palaeoclimatological research has been conducted in the Alp region of Central Europe, aimed at different time perspectives but few useful quantitative temperature reconstructions have so far been produced for the last one or two millennia, although several are under development in the Millennium project founded by the European Union (Gagen et al., 2006). Several of the records from this region that are now available are less suited for being used in a quantitative calibrated temperature reconstruction. The siliceous algae-based Oberer Landschitzsee temperature reconstructions (Schmidt et al., 2007) have a crude resolution and likely contain much non-temperature related information. The Lake Anterne sediment record (Millet et al., 2009) is influenced by anthropogenic activities during the recent centuries, which makes it difficult to compare the level of the medieval warmth to the modern one. The temperature reconstruction from Central Europe by Glaser and Riemann (2009) based on historical documentary sources first starts in the early 1000s and probably underestimates the low-frequency variability in medieval times when the data coverage is sparse. The Spannagel Cave speleothem $\delta^{18}O$ record (Mangini et al., 2005) would be a potentially good record but it unfortunately ends already in the 1930s and thus has too short a calibration period for being really useful. This leaves us with only two records, both reflecting summer temperatures, 1) the varved Lake Silvaplana chironomid-based record

(Larocque et al., 2009), and the tree-ring maximum latewood density record from the Alps by Büntgen et al. (2006).
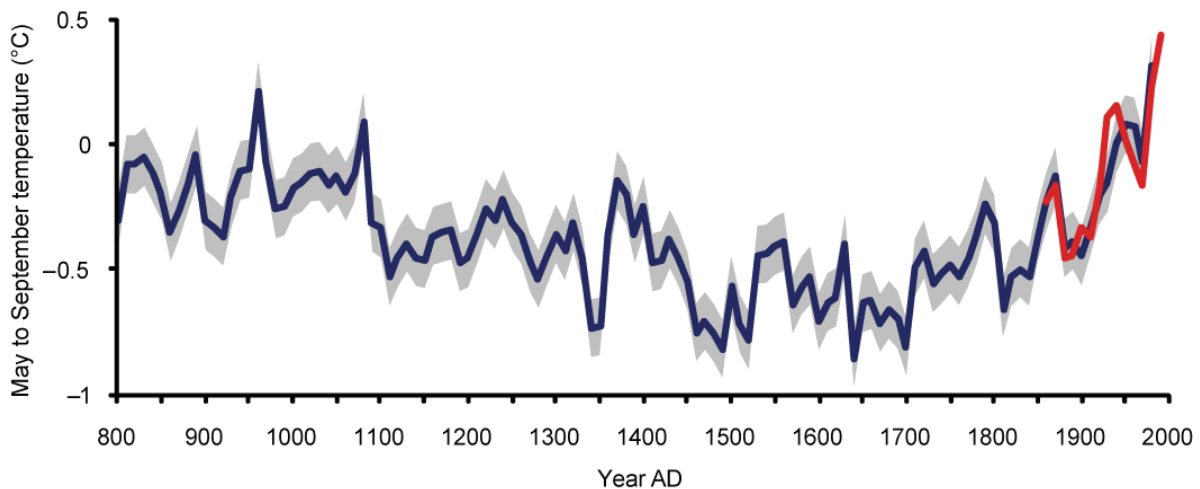

Fig. 9. May to September temperature reconstruction for Central Europe (blue line) calibrated to instrumental temperatures for the Alp region (red line) with 2 standard deviation error bars (grey shading).
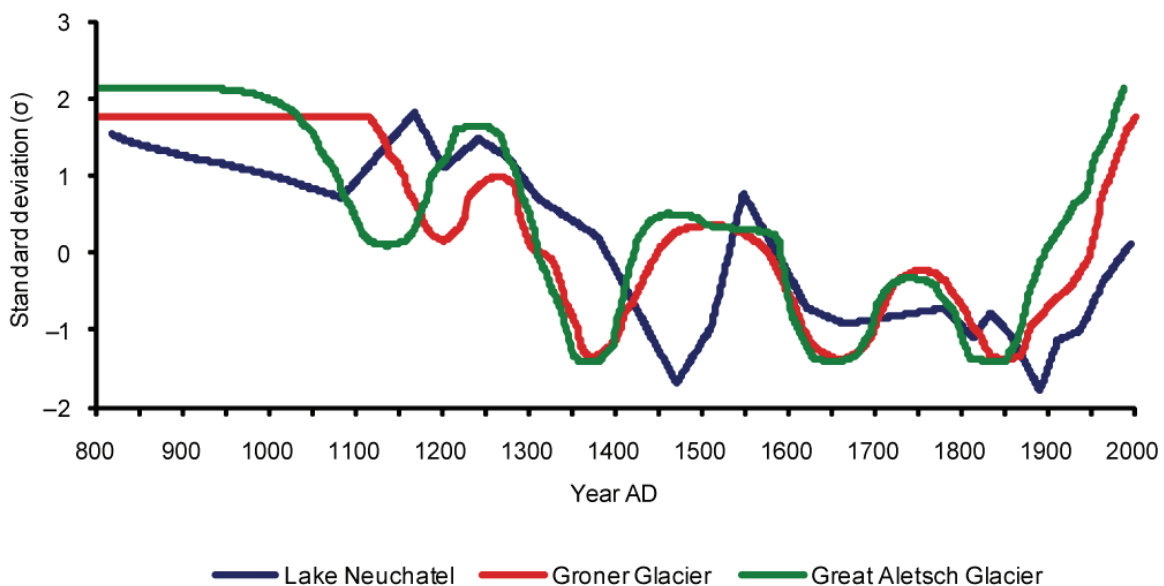

Lake Neuchatel   Groner Glacier   Great Aletsch Glacier

Fig. 10. Three normalised temperature records from Central Europe shown relative to their AD 1000–1899 mean.

Long, homogenised instrumental records are available from Central Europe, going back to ~1760 (Auer et al., 2007; Böhm et al., 2010). Even if we thus have longer instrumental temperature records from Central Europe, we will, for a matter of consistency with the other regional reconstructions, use the May to September 50–45°N × 5–15°E CRUTEM3 grid cells. The correlation coefficient over the calibration period 1850–1999 amounts to 0.84 (r^2 = 0.70). During most of the past 12 centuries, the May to September temperatures of Central Europe have been below the 1961–1990 reference level with the exception for parts of the MWP. In the 10th century, and partly in the 12th century, temperatures seem generally to have been at

or above this reference level. A very clear, albeit somewhat variable, LIA is seen from c. AD 1250 until the mid-19th century with a maximum cooling in the 17th century.

The reconstructed decadal temperature variability of the last 12 decades is about 2°C, with probably the warmest temperatures observed at the end of the 20th century, although this is within the uncertainty level of the medieval warmth (Fig. 9). A medieval warming c. AD 800–1280 is seen, peaking approximately AD 1000, but is interrupted by at least three major cool events: c. AD 1050 and c. AD 1170, and c. 1250. A reconstruction based on only two records cannot be regarded as particularly robust but the low-frequency trend of the reconstruction agrees very well with the with three low-resoled temperature reconstructions from the Alp region (Fig. 10): the glacier length records from the Great Aletsch Glacier and Groner Glacier (Holzhauser et al., 2005) and the pollen-based summer temperature reconstruction from Lake Neuchatel (Filippi et al., 1999) (Fig. 10).

2.5 China

Whereas the coverage of proxy data is very sparse for most of Asia, the coverage for China is rather good. In fact, China is, together with parts of North America, probably the region outside of Europe, with the best palaeoclimatological and palaeoenvironmental records. Previously, much of the research has been published in Chinese, and thus not been easily accessible to Western scholars, but in recent years most Chinese research has been published in English and consequently well incorporated in the international field of palaeoclimatology. The problem remains, however, that relatively few of the reconstructions are available in digital form from public databases. China has the world's longest continuous written historical records that can be used to reconstruct past climate variability. Therefore, much valuable research in the field of historical climatology has been done in China. Many syntheses on past climate in China, using different natural and/or historical archives, have been written and two of them are of relevance to us. Yang et al. (2002) published the first highly resolved annual mean quantitative temperature reconstruction for China covering the last two millennia and Ge et al. (2010) updated this reconstruction and also made reconstructions for different parts of China.
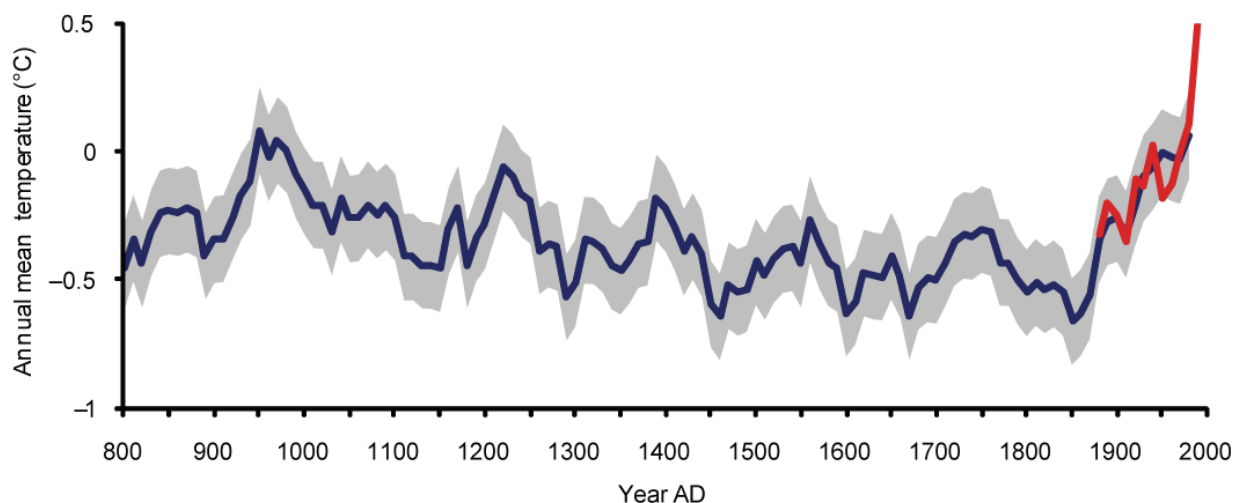

Fig. 11. Annual mean temperature reconstruction for China (blue line) calibrated to instrumental temperatures for China (red line) with 2 standard deviation error bars (grey shading).
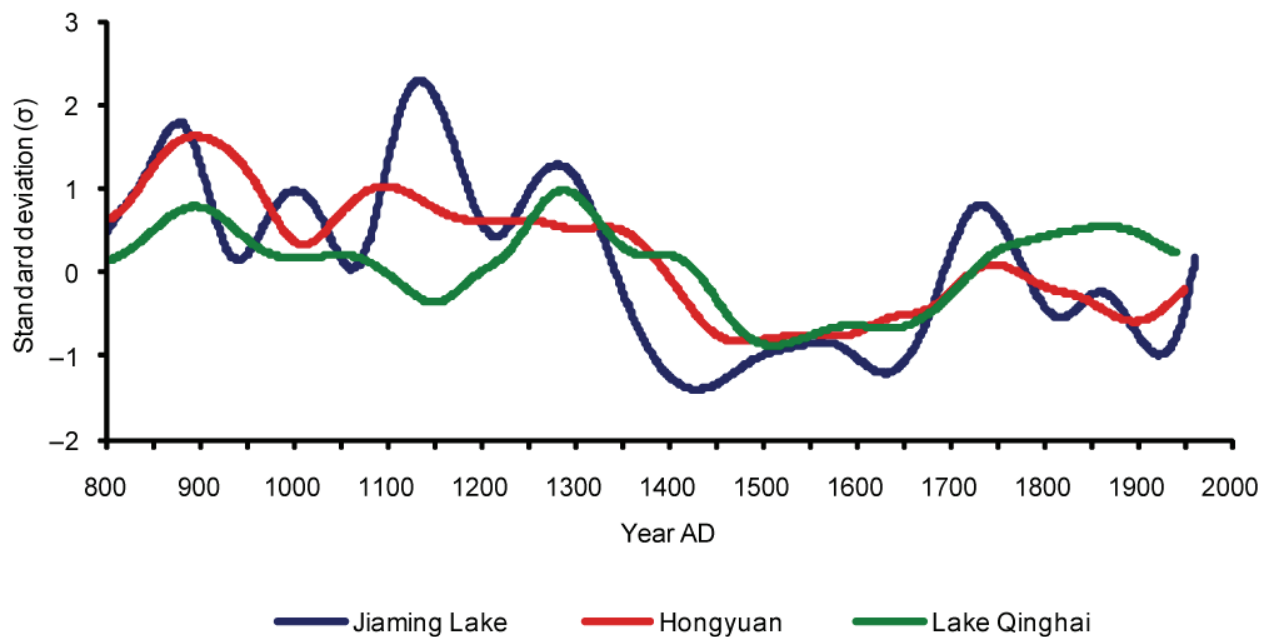
Fig. 12. Three normalised temperature records from China shown relative to their AD 1000–1899 mean. The records have been smoothed with a 100-year cubic spline filter.

Eight records were deemed suitable for reconstructing decadal China temperature variability of the last 12 centuries: 1) the Mongolia tree-ring width record (D'Arrigo et al., 2001), 2) the ShiHua Cave speleothem δ18O record (Tan et al., 2003), 3) a composite of East China documentary records of annual mean temperature (Yang et al., 2002), 4) a composite of documentary East China records of cold season temperature (Ge et al., 2003), 5) the Dulan, northeast Qinghai-Tibetan Plateau, tree-ring width record (Zhang et al., 2003), 6) the Wanxiang Cave speleothem δ18O record (Zhang et al., 2008b), 7) a composite of Zangtze Delta region documentary records of annual mean temperature (Zhang et al., 2008a), and 8) the Heshang Cave speleothem δ18O record (Hu et al., 2008).

The China annual temperature reconstruction has been calibrated against the annual mean temperature from the 55–20°N × 70–135°E CRUTEM3 grid cells. Unfortunately, the meteorological network of instrumental temperature measurements is much shorter and sparser back in time in China than in Europe or most of North America. We can therefore not start the calibration period of the Chinese temperature reconstruction until AD 1880 and the lack of proxy data in the 1990s forces us to end the calibration period AD 1989. Thus, our calibration period for China is just 11 decades. The correlation coefficient over the calibration period amounts to 0.83 (r^2 = 0.68) and thus shows, although the degrees of freedom are limited, a strong relationship between the proxy composite and the instrumental temperatures.

The multi-decadal annual mean temperature variability in China seems to have been slightly less than 1°C and thus smaller then in most other regions (Fig. 11). During most of the past 12 centuries, the temperatures have been well below the 1961–1990 reference level. However, during the second half of the 10th century temperatures equalled or exceeded the 1961–1990 mean. This warm event probably represents the peak of the MWP in China. Warm conditions, similar to those of the 20th century, are also seen in the first half of the 13th century. Five major cooling events during the LIA are visible, centred c. AD 1300, AD 1450, AD 1600, AD 1675 and AD 1850. According to the instrumental record, late 20th century

temperatures may be the highest in the last 12 centuries but this is not seen in the proxy reconstruction itself.

Overall, both the shape and amplitude of the low-frequency temperature variability are well in line with the pollen-inferred temperature reconstructions by He et al. (2004). There also exist some other quantitative temperature reconstructions with lower temporal resolution that can be used to compare and verify the reconstructed low-frequency trend. We have chosen to use the Jiaming Lake (Yang et al., 2002), the Hongyuan Lake (Yang et al., 2002), and the Lake Qinghai (Liu et al., 2006) sediment records for this purpose. They show a quite similar picture of China low-frequency temperature variability, especially the existence of a MWP c. AD 850–1300 and a clear LIA cooling c. AD 1400–1700 (Fig. 12).

2.6 North America

Relatively much data, supposedly reflecting past temperature variability, are available from North America but most of this data consist of tree-ring records. A significant problem is that many of the tree-rings records, especially those from lower latitudes, apparently are more sensitive to precipitation than temperature. All tree-ring width records from high-elevation semi-arid sites in the North American Southwest can be said to possess an ambiguous temperature signal at best, although they have indeed been used in many previous multi-proxy reconstructions (e.g., Cook et al., 2004; Crowley and Lowery, 2000; Esper et al., 2002; Jones and Mann, 2004; Mann et al., 1999; Mann et al., 2008; Mann et al., 2009; Mann and Jones, 2003; Osborn and Briffa, 2006). The potential problems with tree-ring records from semi-arid, high-elevation regions in the North American Southwest have been highlighted in D'Arrigo et al. (2006) and Loehle (2009). Moreover, most of those chronologies have a low replication value back in medieval times. Excluding the tree-ring width records from lower latitudes leaves us with a relatively limited number of palaeotemperature proxy records with adequate resolution and time control to use in a calibrated temperature reconstruction.
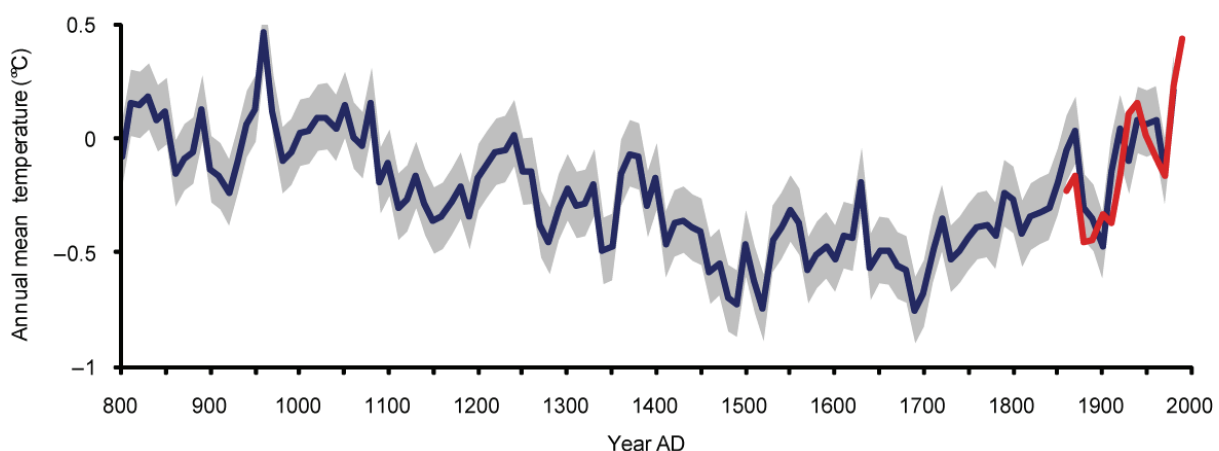

Fig. 13. Annual mean temperature reconstruction for North America (blue line) calibrated to instrumental temperatures for this continent (red line) with 2 standard deviation error bars (grey shading).

Eight records were deemed suitable for reconstructing decadal North American temperature variability of the last 12 centuries: 1) the Devon Island, Canadian Arctic

Archipelago, ice-core δ18O record (Fisher et al., 1983), 2) the Gulf of Alaska tree-ring width record (D'Arrigo et al., 2006), 3) the Iceberg Lake varved sediment record (Loso, 2009), 4) the Canadian Rockies tree-ring width record (Luckman and Wilson, 2005), 5) the Chesapeake Bay sea sediment record (Cronin et al., 2003), 6) the Bermuda Rise sea sediment record (Keigwin, 1996), 7) the Punta Laguna lake sediment record (Curtis et al., 1996), and 8) the Nicoa Cave speleothem δ18O record (Mann et al., 2008). The North America annual temperature reconstruction has been calibrated against the annual mean temperature from the 85–5°N × 50–170°W CRUTEM3 grid cells. The correlation coefficient over the calibration period 1860–1989 amounts to 0.78 (r^2 = 0.61).
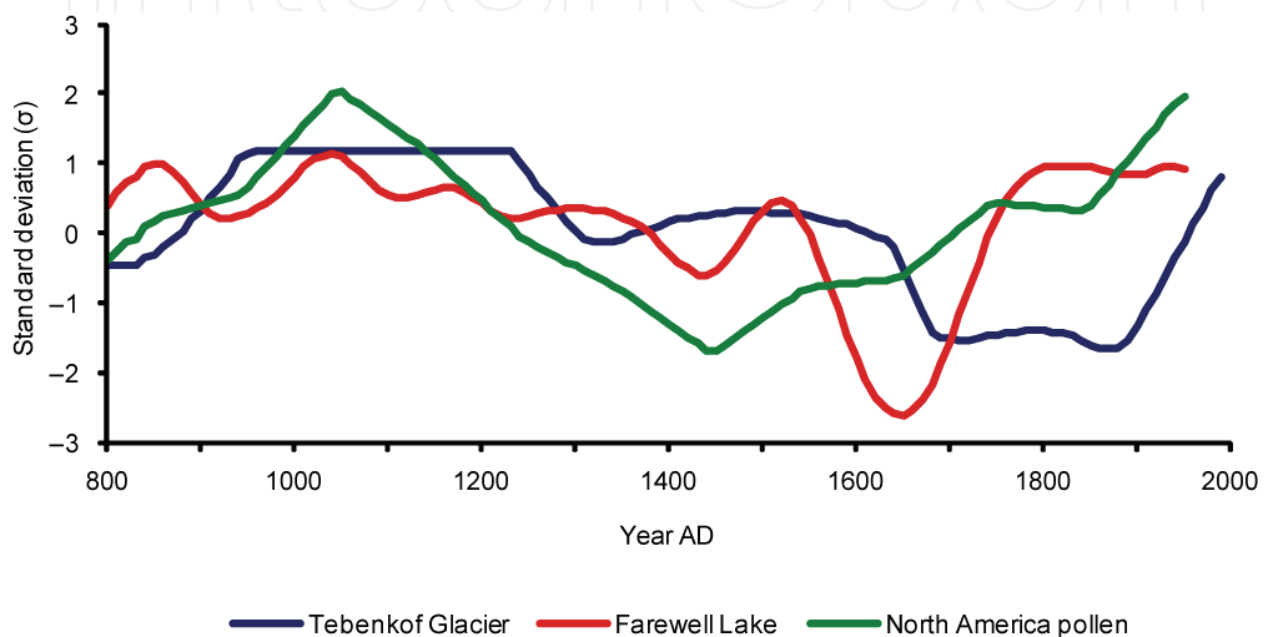

Fig. 14. Three normalised temperature records from North America shown relative to their AD 1000–1899 mean. The Lake Farwell record has been smoothed with a 100-year cubic spline filter.

The reconstructions show that annual mean temperatures have been below the 1961–1990 reference level prior to the 20th century except during the MWP (Fig. 13). A very sharp warm peak during the MWP occur c. AD 960 with temperature levels equalling those of the last two decades. A general cooling, consistent with the LIA, is seen in the North American reconstruction c. AD 1400–1850 with a markedly cold spell also c. AD 1900. The maximum cooling occurs around AD 1500 and AD 1700. The reconstructed decadal variability of the last 12 centuries is slightly more than 1°C. The reconstructed low-frequency trend could be compared to a number of different proxy records with lower temporal resolution. We have chosen to use three that, admittedly, primarily reflect warm season temperatures: The Tebenkof Glacier length record from southern Alaska (Barclay et al., 2009), the Farewell Lake sediment record from central Alaska (Hu et al., 2001), and the entire North American pollen-based temperature reconstruction by Viau et al., 2006. The reconstructed long-term changes are well reflected in the low-resolution records (Fig. 14). The existence of a distinct MWP in North America is verified and also the existence of a LIA, although the three low-resolution records are less consistent when it was coldest.

3. Discussion

The North Atlantic region – from Greenland to Europe – has the best data coverage together with China and the coastal areas of North America. Data are lacking to make meaningful regional reconstructions for Africa, the Middle East or the interior of Asia as well as for all of the Southern Hemisphere (except for, perhaps, South America, see Neukom et al., 2010). Even in relative data rich regions (e.g., Central Europe and Scandinavia) the seasonal bias in the data limits our possibility to make annual mean temperature reconstructions. Most high latitude data, except for the Greenland borehole and $\delta^{18}O$ ice-core records, have a clear bias towards warm season temperatures. We have been able to present regional reconstructions for 1) warm season temperatures of Scandinavia north of 60°N, 2) warm season temperatures for northern Siberia, 3) annual mean temperatures for Greenland, 4) warm season temperatures for the Alp region of Central Europe, 5) annual mean temperatures for China, and 6) annual mean temperatures for the whole of the North American continent.

In all the six reconstructions we can see evidence for the MWP, the LIA and the modern warming. Moreover, in all the six regional reconstructions the medieval warming is peaking in the 10th century. Maximum LIA cooling, however, seems to occur at somewhat different times in the different reconstructions – from the 16th to the 19th centuries – but the 17th century is very cold in all reconstructions. All reconstructions, except the one from central Europe, also agree that the mid to late 19th century was unusually cold, which is of special interest since it is during this period that widespread instrumental temperature measurements start. Late 20th century temperatures are lower than those of the MWP in the Scandinavia, Siberia and especially in the Greenland temperature reconstruction and equal to the medieval warming in the North America reconstruction. However, in the China and the Central Europe reconstructions late 20th century warming exceeds the medieval one, although this is not clear from the proxy reconstructions themselves but only from the instrumental temperature data spliced to the reconstructions.

The trends visible in palaeotemperature proxy records with low (multi-decadal to centennial) resolution, which thus cannot be robustly calibrated to the instrumental record, are generally in good agreement with the trends of the calibrated reconstructions from the same regions. This implies that the calibrated regional reconstructions preserve a good degree of the low-frequency variability although the amplitude of the reconstructed changes probably is too small due to the statistical problems of calibrating noisy data to instrumental data (see, e.g., von Storch et al., 2004; Lee et al., 2008).

Although, from the regional reconstructions presented here, it seems doubtful if the late 20th century warming exceeded the medieval one in the Northern Hemisphere, much more data are needed to draw firm conclusions in this matter. In order to truly assess the spatio-temporal pattern of past temperature variability we also need to be able to make reconstructions from other regions as Africa, the Middle East and Central Asia. Much more research in the field is therefore needed to develop new proxy records from different natural recorders of climate variability, such as fossil pollen, ice-cores, lake and marine sediments, speleothems, and tree-ring width and density, as well as from historical records. It is also very important that all new palaeotemperature records are archived, and thus easily accessible, in public databases as the World Data Center for Paleoclimatology (http://www.ncdc.noaa.gov/paleo/paleo.html).

Although it is outside the scope of this article, we can briefly discuss the possible influence of variations in solar and volcanic forcing on the reconstructed temperatures. All the six

regional temperature reconstructions show some agreement with the assumed low-frequency variability in solar forcing of the last 12 centuries (Bard et al., 2000). The medieval period, with high temperatures, had a general high solar activity, whereas the cold LIA was dominated by lower solar activity (Ammann et al., 2007). The warming in the 20th century coincides with an increase in solar forcing, although the warming trend has probably also been amplified in the last decades by anthropogenic greenhouse gas emissions (IPCC, 2007).

4. Conclusion

The presently available palaeotemperature proxy data records do not support the assumption that late 20th century temperatures exceeded those of the MWP in most regions, although it is clear that the temperatures of the last few decades exceed those of any multi-decadal period in the last 700–800 years. Previous conclusions (e.g., IPCC, 2007) in the opposite direction have either been based on too few proxy records or been based on instrumental temperatures spliced to the proxy reconstructions. It is also clear that temperature changes, on centennial time-scales, occurred rather coherently in all the investigated regions – Scandinavia, Siberia, Greenland, Central Europe, China, and North America – with data coverage to enable regional reconstructions. Large-scale patterns as the MWP, the LIA and the 20th century warming occur quite coherently in all the regional reconstructions presented here but both their relative and absolute amplitude are not always the same. Exceptional warming in the 10th century is seen in all six regional reconstructions. Assumptions that, in particular, the MWP was restricted to the North Atlantic region can be rejected. Generally, temperature changes during the past 12 centuries in the high latitudes are larger than those in the lower latitudes and changes in annual temperatures also seem to be larger than those of warm season temperatures. In order to truly assess the possible global or hemispheric significance of the observed pattern, we need much more data. The unevenly distributed palaeotemperature data coverage still seriously restricts our possibility to set the observed 20th century warming in a global long-term perspective and investigate the relative importance of natural and anthropogenic forcings behind the modern warming.

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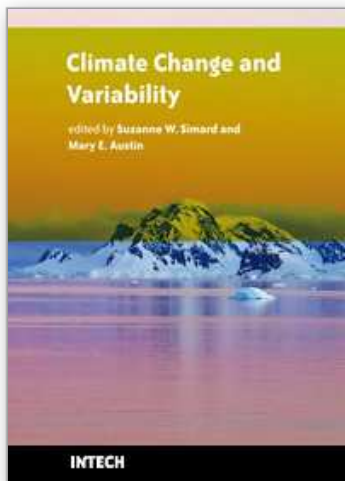
Climate Change and Variability

Edited by Suzanne Simard

Climate change is emerging as one of the most important issues of our time, with the potential to cause profound cascading effects on ecosystems and society. However, these effects are poorly understood and our projections for climate change trends and effects have thus far proven to be inaccurate. In this collection of 24 chapters, we present a cross-section of some of the most challenging issues related to oceans, lakes, forests, and agricultural systems under a changing climate. The authors present evidence for changes and variability in climatic and atmospheric conditions, investigate some the impacts that climate change is having on the Earth's ecological and social systems, and provide novel ideas, advances and applications for mitigation and adaptation of our socio-ecological systems to climate change. Difficult questions are asked. What have been some of the impacts of climate change on our natural and managed ecosystems? How do we manage for resilient socio-ecological systems? How do we predict the future? What are relevant climatic change and management scenarios? How can we shape management regimes to increase our adaptive capacity to climate change? These themes are visited across broad spatial and temporal scales, touch on important and relevant ecological patterns and processes, and represent broad geographic regions, from the tropics, to temperate and boreal regions, to the Arctic.

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