Thousand years of winter surface air temperature variations in Svalbard and northern Norway reconstructed from ice-core data

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Abstract

Two isotopic ice core records from western Svalbard are calibrated to reconstruct more than 1000 years of past winter surface air temperature variations in Longyearbyen, Svalbard, and Vardø, northern Norway. Analysis of the derived reconstructions suggests that the climate evolution of the last millennium in these study areas comprises three major sub-periods. The cooling stage in Svalbard (ca. 800–1800) is characterized by a progressive winter cooling of approximately 0.9 °C century^{-1} (0.3 °C century^{-1} for Vardø) and a lack of distinct signs of abrupt climate transitions. This makes it difficult to associate the onset of the Little Ice Age in Svalbard with any particular time period. During the 1800s, which according to our results was the coldest century in Svalbard, the winter cooling associated with the Little Ice Age was on the order of 4 °C (1.3 °C for Vardø) compared to the 1900s. The rapid warming that commenced at the beginning of the 20th century was accompanied by a parallel decline in sea-ice extent in the study area. However, both the reconstructed winter temperatures as well as indirect indicators of summer temperatures suggest the Medieval period before the 1200s was at least as warm as at the end of the 1990s in Svalbard.

The recent rapid climate and environmental changes in the Arctic, for instance, sea-ice retreat (e.g., Comiso et al. 2008) and ice-sheet melting (e.g., van den Broeke et al. 2009), require a focus on long-term variability in this area in order to view these recent changes in the long-term context. Svalbard is an archipelago in the North Atlantic sector of the Arctic Ocean (Fig. 1). Generally speaking, Svalbard's climate is influenced by sea-surface temperatures in the Norwegian Sea, by sea-ice extent in the Barents Sea and Fram Strait and by the tracks of Arctic weather systems (e.g., Hisdal 1998). Svalbard has a mild and relatively wet climate for its high latitude and, in addition, temperature and precipitation are highly variable during the year due to interaction between the different atmospheric and oceanic systems.

The homogenized Svalbard Airport temperature record starting in 1911 is one of only a few long-term (>65 yr) instrumental temperature series from the High Arctic (Nordli et al. 1996; Nordli & Kohler 2003). Prior to 1912 very few instrumental measurements are available in the Arctic as far north as Svalbard, making the Svalbard Airport temperature record important for interpreting present Arctic climatic trends. Nevertheless, the instrumental period is short.

Ice cores are among the best archives of past climatic and environmental changes. Oxygen isotopes are the

Fig. 1 Map of the study area showing the locations referred to in the text.

most commonly used temperature proxy from ice cores (Dansgaard 1964) despite complications around the transfer function relating snow oxygen isotope content to local temperature. A good example is the work by Schneider & Noone (2007), who used $\delta^{18}O$ records from over 40 ice cores from a variety of geographical areas and found that, generally, these records show a large similarity to the global temperature trend over the instrumental record period (1860–1989).

During the past three decades, several ice cores have been drilled in Svalbard mainly by groups from the former Soviet Union and Japan. For a general overview of research on the Svalbard ice cores we refer to Tarrusov (1992), Watanabe (1996), Kotlyakov et al. (2004) and Motoyama et al. (2008). Most of these cores were dated using the accumulation rate deduced from the depth of the 1961–63 radioactive layer. Analysis of Svalbard and other lower elevation glaciers ice cores have shown that alteration of primary chemistry signals by summer surface melt is not so severe as to obliterate seasonal variations (e.g., Koerner 1997; Pohjola, Moore et al. 2002).

Two of the most recent western Svalbard ice cores were drilled at Lomonosovfonna in 1997 (Isaksson et al. 2001) and at Holtedahlfonna in 2005 (Sjögren et al. 2007). The core from the summit of Lomonosovfonna at 1250 m a.s.l. is to date the most comprehensively studied of all the Svalbard ice cores. In this paper we use the $\delta^{18}O$ records from these two Svalbard ice cores to reconstruct the surface air temperatures (SAT) for Longyearbyen and Vardø, in northern Norway. To our knowledge this is the first attempt to date to quantify in detail past temperature changes in Svalbard from isotopic ice-core records. An alternative approach based on inversion of the Lomonosovfonna borehole temperature profile has been successfully applied by van de Wal et al. (2002). This effort yielded estimates of the recent secular SAT trend in the area that extended beyond the available instrumental temperature data. Utilization of the down-core isotopic profile, in comparison, can potentially provide access to a much higher temporal resolution for the reconstruction. The prospective time span that can be covered is comparable to the length of the core in the time domain

and constrained only by the quality of the dating of the analysed ice core.

Data sets

Ice core data

For the temperature reconstruction for Svalbard we have used the ice cores from Lomonsovfonna and Holtedahl-fonna as these provide the best resolution oxygen isotope records suitable for a SAT reconstruction. For the Lomonosovfonna ice core there is also deuterium excess data available.

Lomonosovfonna. This 121-m deep ice core was drilled at the summit of the ice field Lomonosovfonna (at 78° 51′N, 17° 25′E, 1250 m a.s.l.; Fig. 1) in the spring of 1997. The total ice depth at the drill site was estimated to be just a few metres more than the ice-core length. General information about the ice core and analyses can be found in Isaksson et al. (2001). The core dating is based on the updated (Kekonen, Moore, Perämäki, Mulvaney et al. 2005) chronology using the well-known Nye (1963) relation, constrained by depths of the radio-activity peaks found in the core, which appear in 1962–63 and 1954 (Pinglot et al. 1999), the 1903 Grímsovötn (Wastegård & Davies 2009) and the 1783 Laki (Kekonen, Moore, Perämäki & Martma 2005) volcanic eruptions detected by ion analyses and the identification of tephra particles. The annual accumulation rate at the core site is about 0.36 m w. eq. Most of the core was sampled for $\delta^{18}O$ at 5-cm resolution (Isaksson et al. 2001). In total 2840 oxygen isotope samples were analysed, including 543 additional new samples in the 74 to 90 -m section of the core sampled at 2.5-cm resolution. This made it possible to achieve sub-annual time resolution as far back as approximately 1456 AD, with as many as 23 samples per year at the core top. This higher resolution oxygen isotope stratigraphy has also been used to extend the summer peak counting back to 1613 AD, providing an annual timescale for this time period. This is some 100 years longer than in the previous version of the chronology (Pohjola, Martma et al. 2002; Kekonen, Moore, Perämäki, Mulvaney et al. 2005). The timescale error for the section of the core younger than 1613 AD is estimated to be ± 1 year in the vicinity of the tiepoints used and roughly ± 5 years between the dating horizons. The latter value was estimated by a comparison of the sulphate record in the core with the dates of known volcanic eruptions (Moore et al. 2006).

The chronology of the deeper section below 90.61-m core depth (76.2 m w. eq., before 1613 AD) is based on Nye age modelling (Nye 1963) with the accumulation rate of 0.36 m w. eq. year^{-1}—the average value for the 1783–1997 period. Modelling results suggest the ice in the very bottom of the core at 121.6-m core depth (103.3 m w. eq.) is dated to 770 ± 150 AD, which is much older than the previous estimate of about 1100 AD. Since the Nye model-based dating below the three-quarter depth to the bedrock is often considered less reliable, a reasonable estimate of the timescale error and reconciliation with the potential time markers are required. The timescale error in the core chronology below 90.61 m is estimated as

$$\Delta_t = \sqrt{\Delta_H^2 + \Delta_\lambda^2 + \Delta_{tc}^2},$$

where Δ_H, Δ_λ and Δ_{tc} are the mutually independent errors due to uncertainties in the ice cap thickness (2.0 m), mean accumulation (0.06 m w. eq. year^{-1}) and the oldest counted date (5 years). The actual errors are the functions of depth and calculated by taking the derivative on the correspondent variable in the Nye formula. Fig. 2 depicts the updated ice-core chronology including the major time markers.
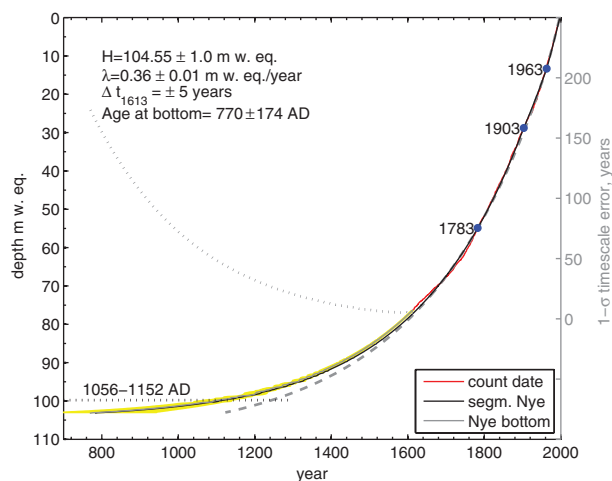

Fig. 2 Updated timescale of the Lomonosovfonna core based on two volcanic markers from the 1783 Laki eruption and the 1903 Grímsovötn eruption, the 1963 tritium peak (blue circles), layer counting back to 1613 (red line) and Nye modelling below this date (grey line). Solid black line shows the Nye model of the entire core chronology estimated from mean accumulation rates of 0.42, 0.32 and 0.37 m w. eq. between the three dated segments (1997–1963–1903–1783) and 0.36 m w. eq. year^{-1} below the 1783 horizon. Yellow highlights the ± 1 standard error in the timescale below the oldest counted date of 1613 AD; also shown as a dotted grey line on the right axis. The dashed grey line shows the previous timescale of the ice core.

Comparing the sulphate record in the core with the history of volcanic eruptions suggests that the pronounced peak at 117.6-m core depth (99.8 m w. eq.; Moore et al. 2006) dated to 1103 ± 90 AD is more likely to be ascribed to the strong eruptions of the Iceland's Hekla volcano in the years 1104 and 1158 (Wastegård & Davies 2009). Both events formed tephra layers in a number of locations in north-western Europe, including coastal northern Norway, implying favourable circulation patterns for the transport of ash and aerosols to Svalbard. The previous association of this peak with a much stronger but more remote eruption of an unknown volcano in the year 1259 (Moore et al. 2006) is now cumbersome, as reaching this horizon would require a constant accumulation rate of about 0.6 m w. eq. year^{-1} from 1613 back in time. A secondary peak in the sulphate record at 114.2-m core depth (96.8 m w. eq.) that has not reached the significance threshold in Moore et al. (2006) and dated to 1245 ± 62 AD in the updated chronology can be related to the 1259 eruption instead. The precise attribution is, however, hampered by discontinuities in the chemical stratigraphy in the bottom section of the core, with the two most prominent gaps in the 112.5–113.8 and 115.2–116.4-m core depth intervals. They cover the time intervals of 1162–1208 ± 70 AD and 1255–1292 ± 55 AD and could potentially accommodate the 1259 sulfate peak as well. In view of the additional uncertainties associated with applicability of the Nye model at the depths close to the bedrock, as well as unknown past dynamics of the ice cap, the results for the bottom section of the core (before ca. 1300 AD) should be interpreted with caution.

Figure 3 shows the raw isotopic series together with the 11-year running mean to highlight the decadal scale variations in $\delta^{18}O$ put on the updated Lomonosovfonna chronology. Various aspects of the oxygen isotope 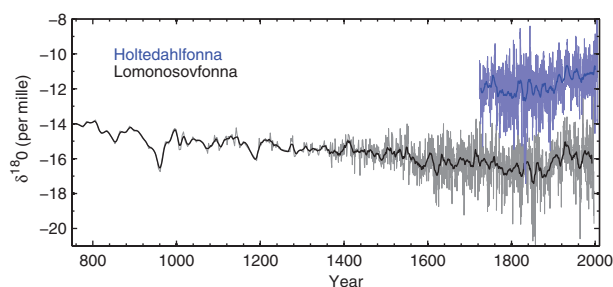record from the Lomonosovfonna ice core have previously been discussed in papers by Isaksson and co-workers (Isaksson et al. 2001, 2003; Isaksson, Divine et al. 2005; Isaksson, Kohler et al. 2005) and Fauria et al. (2009). Deuterium excess records from Lomonosovfonna span the period 1400–1997 and have been described and discussed in Divine et al. (2008).

Holtedahlfonna. In April 2005 one ice core was drilled to a depth of 125 m at Holtedahlfonna (79.13° N, 13.27° E at 1150 m a.s.l.) about 40 km north-east of Ny-Ålesund, on the west coast of Spitsbergen (Fig. 1). Radar measurements showed that the ice depth in the area is highly variable and that it is approximately 150–175 m at the drill site. However, due to water-saturated ice layers in the area, the exact depth from radar surveys is uncertain and possibly considerably deeper. The core segments were packed in plastic bags in the field and kept in insulated boxes at temperatures below freezing during transport to cold room facilities at the Norwegian Polar Institute in Tromsø, Norway. The estimated age of this core is determined by glaciological modelling to be about 300 years for an ice depth of 150 m (Sjögren et al. 2007).

Samples for oxygen isotopes were cut at 5-cm resolution in the uppermost 55 m and 2.5 cm below that. This gives between 14 and 6 samples/yr to the bottom of the core. Altogether 3500 samples were analysed. The upper part of Holtedahlfonna was measured using a Delta E mass spectrometer (Thermo Fisher Scientific, Waltham, MA, USA), whereas the lower part of Holtedahlfonna ice core was analysed using a Delta V Advantage mass spectrometer (Thermo Fisher Scientific) with a GasBench II (Thermo Fisher Scientific), both at the Institute of Geology, Tallinn University of Technology.

Tritium analyses were performed at the Centre for Isotope Research, University of Groningen, on samples from 5-cm intervals and the 1963 peak level was identified at about 28.4-m core depth. This gives an annual accumulation rate of 0.52 cm w. eq. during the period 1963–2005, which compares favourably with the results from a core drilled at the same site in 1992 (Kameda et al. 1993). This shows that the 2005 core is representative for accumulation conditions in the area. Dating of the core is based on a combination of counting summer peaks in the oxygen isotope stratigraphy (Pohjola, Martma et al. 2002), the tritium date at 1963, the 1783 Laki volcanic eruption and a Nye model with a prescribed ice depth at the core site of 300 m. This depth, when used in glaciological modelling, was found to provide the best fit between the chemical series of the well-dated Lomonosovfonna core and the Holtedahlfonna core (Moore et al. 2010). Subsequent application

Fig. 3 Raw isotope $\delta^{18}O$ records of the Lomonosovfonna (grey) and Holtedahlfonna (light blue) ice cores. Black and blue solid lines show 11-year running means estimated from annually resampled data. Note that for convenience the Holtedahlfonna series is shifted up by three per mille.

of the method described in Moore et al. (2006) to the nine chemical series of the Holtedahlfonna core has revealed a pronounced peak in sulphate record at 88.9 -m depth, which was identified as the Laki 1783 eruption (Moore et al. 2010). In the present study we used the section of the core dated back to 1723 for analysis, shown in Fig. 3.

Instrumental records

As a target for reconstruction we utilize the homogenized Longyearbyen–Svalbard Airport (78.25° N, 15.47° E, 27 m a.s.l.) monthly temperature record, which starts in 1911 (Nordli et al. 1996; Nordli & Kohler 2003) and Vardø (70.54° N, 30.61° E, 12 m a.s.l.) series, which has been now extended back to 1840 (Polyakov et al. 2003). These two sites are located quite far apart and separated by the Barents Sea. However, their temperature anomalies are coherent during the 1912–2009 period of overlap, as confirmed by the correlation coefficients of 0.68 for annual means and 0.54 for December–February (DJF) means.

Data treatment

Prior to further statistical analysis the raw unevenly sampled oxygen isotope series were resampled to the annual scale. This was done by aggregating and averaging all $\delta^{18}O$ values from samples ascribed to the annual interval centred at the first Julian day according to the established depth–age chronology. In order to keep a consistent annual time increment throughout the record we filled the gaps in the data by spline interpolation prior to smoothing. There are 460 interpolated data points in total in the Lomonosovfonna oxygen isotope series. All of them occur prior to 1525 AD, where the annual resolution is not always attained, with 52 interpolated points in the period 1300–1525 AD. For the Holtedahlfonna ice core, the sampling density was sufficient to sustain a sub-annual resolution along the entire core length; no interpolation procedure was therefore required for this oxygen isotope series.

Temperature reconstruction

In reconstructing past changes of an environmental variable (e.g., SAT), proxy time series are typically scaled to, or regressed against, instrumental climate data. Both approaches rely implicitly on the simplifying assumption of a stable in time linear relationship between a predictor proxy and predictand climate series. The term "scaling" refers to the equalization of the mean and standard deviation (STD) of a proxy time series to the corresponding values of an instrumental record over a defined period of overlap (the so-called calibration period).

Given the instrumental and the proxy series, T_{inst} and P, the reconstructed temperature T_{rec} is calculated as

$$T_{rec} = (P - \bar{P}_{cal}) \frac{STD(T_{inst})}{STD(P_{cal})} + \bar{T}_{inst},$$

where $\bar{P}_{cal}$ and $\bar{T}_{inst}$ designate the means of the proxy series and the instrumental temperatures, respectively, during the calibration period.

When a least squares linear regression approach is used for calibration, the resulting amplitude of the reconstruction equals the amplitude of the instrumental data multiplied by the Pearson correlation coefficient α between the proxy and instrumental record during the calibration period (Esper et al. 2005). As a result, the reconstructed amplitude of the environmental variable is reduced by the square root of "unexplained" variance, i.e., $\sqrt{1 - \alpha^2}$, in the regression. The drawback of scaling is inflated reconstruction error variance (Esper et al. 2005).

In proxy-based climate reconstructions a common situation is that the proxy series efficiently captures the longer term variability, yet for various reasons shows no coherency with the target environmental variable(s) at the higher resolution timescales. Isotopic signals in precipitation and, hence, in the ice core are known to be an integrated tracer of the water cycle (e.g., Alley & Cuffey 2001). The actual value of $\delta^{18}O$ in precipitation depends on a number of factors other than the ambient temperature alone, such as, for example, variations in the moisture source region(s) (Boyle 1997) or moisture transport pathways affecting fractionation en route. Bias due to changes in precipitation seasonality can potentially influence the isotopic signal in the ice core at annual and longer timescales (Werner et al. 2001; Krinner & Werner 2003). As a result, the local $\delta^{18}O$–SAT relationship often varies in time. The problem can be alleviated by smoothing; that is, low-pass filtering of the scaled proxy record interpreted in terms of the environmental variable. It also efficiently reduces the effect of the inflated due to scaling reconstruction error variance.

Uncertainties in the elaborated ice-core timescales, together with the effects of other processes causing temporal variations in the $\delta^{18}O$–SAT relationship, decrease the Pearson correlation between the annually resampled series to still statistically significant but essentially low, in terms of reliable reconstruction, values of 0.23 (Longyearbyen) and 0.25 (Vardø). However, the decadal and longer term components of the estimated annual mean $\delta^{18}O$ better reflect the SAT variations (see Table 1). Our choice of the scaling approach to

Table 1 Correlation coefficients between the ice core $\delta^{18}O$ and instrumental surface air temperatures (SAT) series used in this study.

	Lomonosovfonna $\delta^{18}O$ series	Holtedahlfonna $\delta^{18}O$ series	Stacked Lomonosovfonna and Holtedahlfonna $\delta^{18}O$ series
Longyearbyen DJF[a] SAT			
annual mean $\delta^{18}O$	0.23	0.13	0.32
5(11)-year running mean $\delta^{18}O$	0.54 (0.67)	0.32(0.49)	0.58 (0.71)
Vardø DJF SAT			
annual mean $\delta^{18}O$	0.25	0.25	0.35
5(11)-year running mean $\delta^{18}O$	0.56 (0.71)	0.61(0.78)	0.66 (0.81)
Longyearbyen annual mean SAT	0.15	0.23	0.27
Vardø annual mean SAT	0.17	0.33	0.35

[a]December–February.

reconstruct past Svalbard SAT is motivated by the need to adequately reproduce the low frequency components of the variability, for which the coherency with the instrumental data is higher.

Fig. 4 shows seasonal variations in SAT and precipitation calculated from the composite Longyearbyen (Svalbard Airport) series for the shorter period of 1959–2001. A pronounced increase in variability during winter compared to summer SAT points to a higher proportion of variance in the annual mean SAT due to winter temperature variations. This is typical for the (sub)polar regions and has been attributed to a more vigorous extratropical winter atmospheric circulation (e.g., Hurrell 1996). The lack of seasonal variability in the amount of precipitation, at least within the considered period, suggests that the isotopic records are expected to be more representative of winter temperature variations in Svalbard. This assumption is examined through calculation of the correlation coefficients between the annual mean $\delta^{18}O$, 5- and 11-year means and the seasonal and annual mean SAT for Svalbard and Vardø. Results presented in Table 1 confirm there is generally a higher

proportion of common variance between winter temperature variations in Svalbard and the annual mean Lomonosovfonna $\delta^{18}O$ and stacked (see below for details) Lomonosovfonna and Holtedahlfonna $\delta^{18}O$ compared to annual SAT equivalents. This provides strong evidence to support our choice of the winter season SAT as a suitable target for subsequent reconstruction.

Scaling of the Lomonosovfonna $\delta^{18}O$ series against DJF surface air temperatures SAT from Longyearbyen yielded an approximately 1200 year-long SAT reconstruction of Svalbard DJF SAT. The period of 1911–97 used for calibration was the period of overlap between the proxy and the available instrumental data from Longyearbyen. The same procedure was repeated for Vardø temperature series with a calibration length interval of 1850–1997. Fig. 5 displays the results of the analysis. In order to highlight the decadal variations in the reconstructed temperatures we smoothed the series using an 11-year running mean. Two grades of yellow in Fig. 5 show the uncertainty ranges on the reconstructed SAT series estimated for the 5- and 11-year running means.
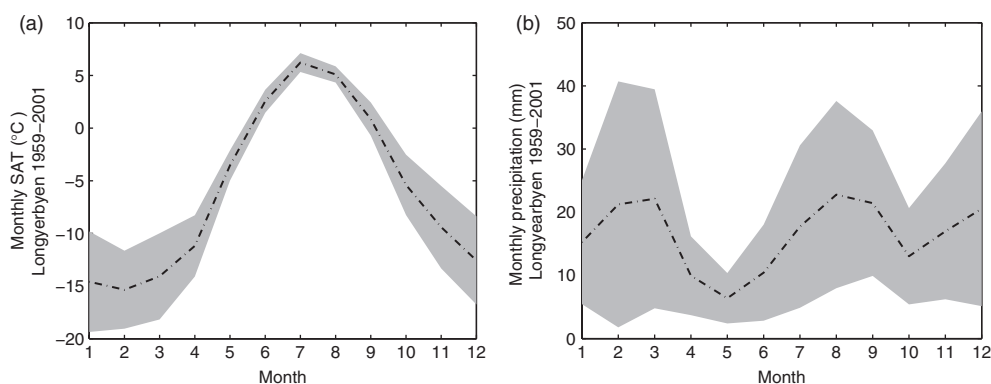

Fig. 4 (a) Monthly mean surface air temperature (SAT) and (b) precipitation in Longyearbyen during 1959–2001 as estimated from the instrumental Svalbard Airport series. Shading highlights the ± 1 STD range.
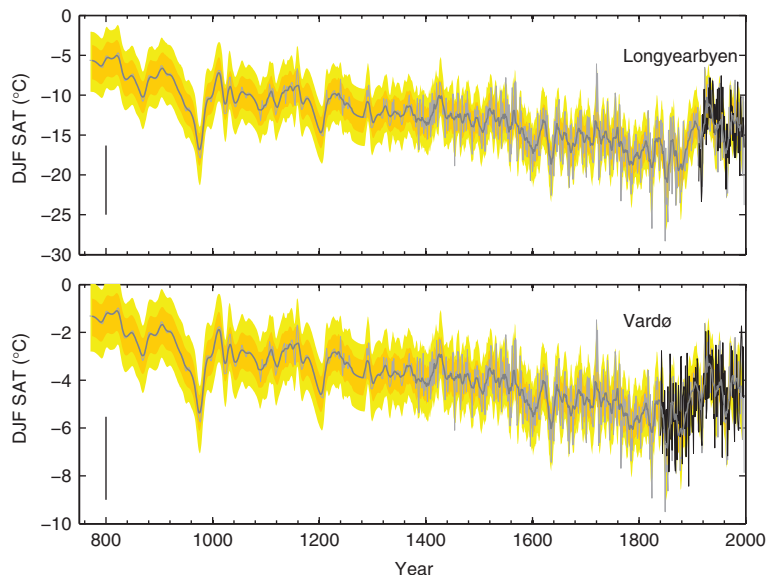
Fig. 5 Longyearbyen and Vardø reconstructed December–February (DJF) surface air temperatures (SAT) from the Lomonosovfonna ice core δ^{18}O series (grey) together with its 11-year running mean (dark grey). The black solid line shows the instrumental data that were used for calibration. Vertical bar shows the double STD of the residual of the reconstruction, with light and dark yellow areas to indicate ± 1 standard error for the 5- and 11-year running means of the reconstructed (predicted) DJF temperature.

The low correlation between the annual mean proxy and instrumental series fundamentally limits the predictive skills of the reconstruction at this timescale. These are, however, improved when the longer scales of the variability are considered. For the smoothed with rectangular filter (i.e., equal filter weights) reconstruction, the standard error is calculated by Briffa et al. (2002; corrected by Gouirand et al. 2008) as

$$[\sigma_{T\text{rec}}(t)]^2 = \frac{1}{\alpha_{\text{rm}}^2}\left(\frac{s^2}{n_{\text{rm}}} + \frac{s^2}{n_{\text{cal}}} + [\check{T}_{\text{rec}}(t) - \check{T}_{\text{rec}}^{\text{cal}}]^2 \sigma_{\text{slope}}^2\right),$$

where s^2 denotes the residual temperature variance, $\check{T}_{\text{rec}}^{\text{cal}}$ is the mean reconstructed temperature over the calibration period of the length n_{cal}, σ_{slope} represents the standard error of the slope estimated from regressing the scaled annual mean δ^{18}O on instrumental DJF SAT and $\check{T}_{\text{rec}}(t)$ is a smoothed reconstructed DJF SAT with n_{rm} being a width of the running mean filter used. The factor of $1/\alpha_{\text{rm}}^2$ where α_{rm} denotes the Pearson correlation coefficient between the smoothed reconstruction and the instrumental series accounts for the use of scaling instead of regression to estimate past SAT (Esper et al. 2005).

In order to reduce uncertainties due to dating errors and other forms of noise in a single oxygen isotope series, we constructed a composite record of the two analysed Holtedahlfonna and Lomonosovfonna δ^{18}O series. The stacked record spans the period of 1723–1997 AD. It was produced by simple averaging of the normalized δ^{18}O

records relative to the common period. The composite records scaled to match the recorded DJF temperature variations on Longyearbyen and Vardø are shown in Fig. 6. Stacking of the two records increased the correlation between the winter temperatures (see Table 1) and reduced the prediction error on the 5-year mean reconstructed temperatures from 4.0 to 3.8 °C on average (1.3 °C vs. 1.0 °C for Vardø). For the 11-year mean DJF SAT estimates, the benefit of combining the two isotopic series was a decrease in error from 2.2 to 2.0 °C for the Longyearbyen reconstruction and from 0.7 to 0.6 °C for the Vardø reconstruction.

Discussion

The results of scaling suggest that the mean DJF winter temperatures vary within a range of –5 to –25 °C for Longyearbyen or –1 to –10 °C for Vardø, corresponding to a span between the most extreme warm (870±140 AD) and cold (1850±3 AD) registered event periods in the record. On decadal timescales and longer, the DJF SAT typically varies within the range of −6 to −17 °C for Longyearbyen and −2 to −6 °C for Vardø. We note that at these timescales the minima of the reconstructed DJF SAT variations correspond reasonably well to the maxima in winter sea-ice expansion to the south of Svalbard, as shown by evidence from historical sea-ice observations from the Greenland (Fig. 7) and Barents (not shown) seas (Vinje 2001; Divine & Dick 2006). This is in line with the
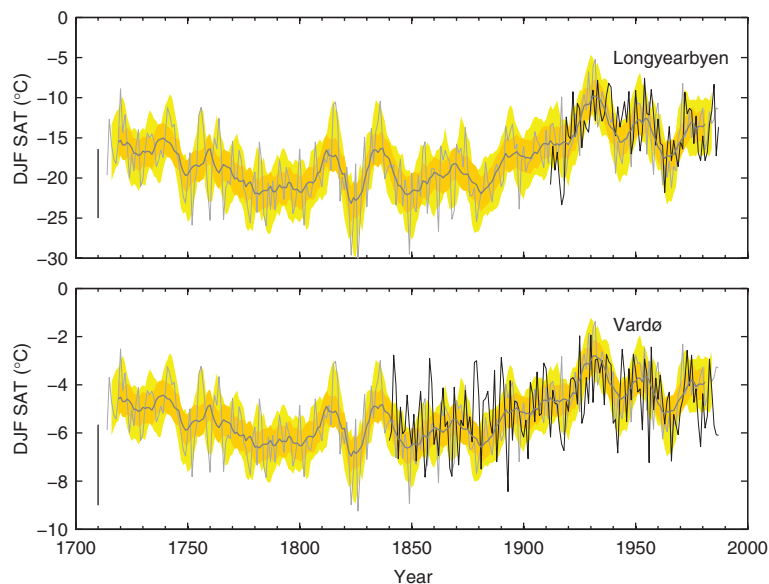
Fig. 6 Longyearbyen and Vardø reconstructed December–February (DJF) surface air temperatures (SAT) from the stacked δ^{18}O Lomonosovfonna and Holtedahlfonna ice core series (grey) together with its 11-year running mean (dark grey). The black solid line shows the instrumental data that were used for calibration. Vertical bar shows the double STD of the residual of the reconstruction with light and dark yellow areas to indicate ± 1 standard error for the 5- and 11-year running means of the reconstructed (predicted) DJF temperature.

inferred SAT–sea-ice link for this area (Benestad et al. 2002), as well as the known effects of sea ice on the distillation history of an air mass and, hence, δ^{18}O in precipitation (Noone & Simmonds 2004). The close connection between the Lomonosovfonna δ^{18}O and sea-ice extent in the study area has recently been used to reconstruct past sea-ice variations in the Nordic seas (Fauria et al. 2009). A reasonably good agreement between sea-ice extent variations in the study area and methanesulfonic acid—another potential ice-core sea-ice

indicator—has been also demonstrated by O'Dwyer et al. (2000).

The climatic tendency revealed by the reconstructions presented here comprises a long progressive winter cooling of a magnitude of approximately 0.9 °C century^{-1} that commenced in Svalbard as early as before the 1000s. This makes it difficult to associate the onset of the Little Ice Age in Svalbard with any particular time period. During the 1800s, which was the coldest period in Svalbard according to the reconstruction, the DJF
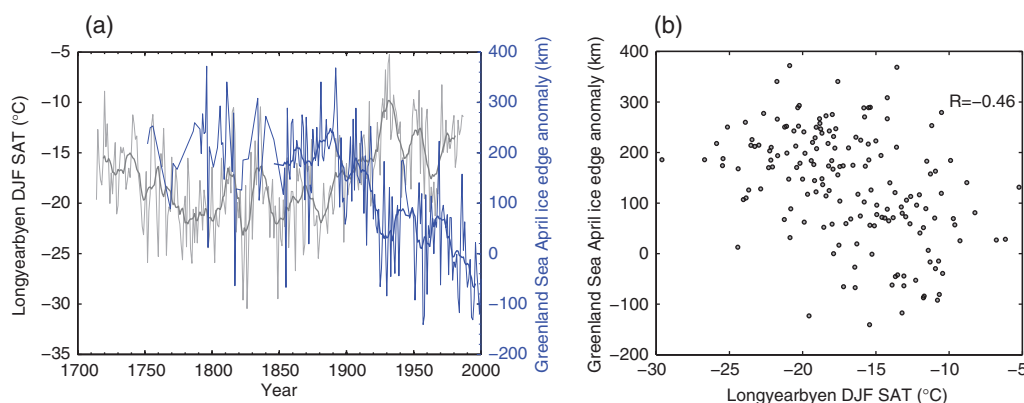

Fig. 7 (a) Longyearbyen reconstructed December–February (DJF) surface air temperatures (SAT) from the stacked δ^{18}O Lomonosovfonna and Holtedahlfonna ice core series (grey) and Greenland Sea April ice extent ice edge anomaly (blue), also shown as (b) a scatter plot. Thick solid lines show the 11-year running means. Due to the sea-ice data sparsity prior to the mid-1800s, the running mean is estimated only for the period of 1850–onwards.

temperatures were as low as -18 °C on average. The corresponding estimated values for the Vardø reconstruction differ, being 0.3 °C century^{-1} for the long-term cooling trend, which culminated in a lasting period with DJF temperatures of the order of -5.5 °C during the 1800s. Comparison of the reconstructed SAT for the two periods of the 1800s and 1900s allows estimation of the Little Ice Age–associated winter cooling, which was on the order of 4 °C and 1.3 °C for Svalbard and northern Norway, respectively. Analysis of the measured temperature profile in the Lomonosovfonna borehole (van de Wal et al. 2002) revealed a warming trend of about 2 °C in the annual mean SAT during the 20th century. We note that despite the fact that trends in the annual mean and winter temperatures are not necessarily directly related, this value lies in fairly good agreement with our δ^{18}O-based estimates of the changes in the corresponding winter SAT.

The inferred ice-core based past SAT variability in Longyerbyen and Vardø agrees very well with available long-term instrumental as well as reconstructed air temperature records from the Arctic region. A gradual increase in reconstructed Svalbard (northern Norway) SAT since the mid-19th century, with an abrupt warming during the 1910s–1920s, is a part of the ongoing well documented pan-Arctic warming (e.g., Polyakov et al. 2003; Bengtsson et al. 2004; Hanna et al. 2004; Box et al. 2009). In a millennial perspective, a pattern of the long-term cooling, which preceded the contemporary warming, has been revealed from the western Arctic ice cores (Kotlyakov et al. 2004) as well as the multiproxy reconstruction of Arctic summer SAT (Kaufman et al. 2009). Kaufman et al. associated a millennial-scale cooling with the slow reduction in summer insolation at high northern latitudes due to changing orbital forcing. Although the cooling trend magnitude in the summer SAT is much lower, on the order of 0.02 °C century^{-1} estimated for the whole Arctic region, one should notice a qualitative similarity between the proxy-inferred temporal evolutions of the air temperature. The different target seasons for reconstruction used, together with a substantial sensitivity of Svalbard climate in winter to sea-ice extent variations in the area, provide a reasonable explanation for this discrepancy.

Additional evidence of the intermittent decoupling of summer and winter temperatures in the study region comes from the analysis of Svalbard ice-core data. Grinsted et al. (2006) calculated the time variations of the amplitude of the annual cycle in the Lomonosovfonna δ^{18}O during the period 1600–1997. This value, which can be considered as a proxy for continentality of the study area, has undergone substantial changes,

implying different temporal evolution of winter and summer SAT in the reference period. The most prominent features are increasing continentality during the 1600s–1700s (onset of the Little Ice Age) followed by a progressive decrease since the culmination of the Little Ice Age in the mid-1800s. A similar increase in the amplitude of the annual temperature cycle in the 19th century is evident from Greenland instrumental data (Box et al. 2009; data available after 1840), Iceland and a number of longer instrumental series from northern Europe including Vardø (e.g., Klingbjer & Moberg 2003; Hanna et al. 2004).

We note that changes in continentality may also have implications for the secular changes in the seasonal distribution of precipitation in Svalbard and hence interpretation of the isotopic signal in terms of past temperature variations. More continental locations tend to receive a higher proportion of their annual precipitation during summer. The annual mean δ^{18}O in Svalbard precipitation can, as a result, be biased towards more isotopically warm values during the Little Ice Age also affecting the inferred air temperatures. Since no precipitation data are available for Vardø prior to the mid-1950s, one can only rely on indirect evidence. In order to explore the potential for this effect we compared the magnitudes of trends in the standardized instrumental Vardø DJF SAT and stacked δ^{18}O series for the common period of 1840–1997. The trend magnitudes appear to be similar within the uncertainty range (indicated in parenthesis), being 0.008(0.002) yr^{-1} and 0.011(0.02) yr^{-1} for the standardized SAT and δ^{18}O series, respectively. In fact, the trend for the isotopic series is even slightly steeper than the corresponding trend in DJF SAT. We note that the opposite would be the case if the analysed isotopic series were substantially affected by the assumed shifts in the seasonality of precipitation. It indicates that the role of this effect, at least within the considered period, is likely negligible.

A long-term cooling revealed in winter temperatures in Svalbard and northern Norway is in line with a multicentennial decline in winter North Atlantic Oscillation that occurred during approximately the same period of time and marked the transition from the Medieval Climate Anomaly to the Little Ice Age in Europe (Trouet et al. 2009). It is known that on annual to sub-decadal timescales the effects of North Atlantic Oscillation–related variability on Svalbard climate are not well pronounced (e.g., Marshall et al. 2001; Zhang et al. 2004). On the longer timescales, however, such a shift to a persistently negative North Atlantic Oscillation would imply a decreasing atmospheric and oceanic heat flux to

the north-east and, hence, a gradual cooling and sea-ice expansion in the Nordic seas (e.g., Marshall et al. 2001).

One of the most striking features of the reconstruction is a lasting pre-1300 period of warm winters during which DJF temperatures were comparable, within error range of those that were observed in Svalbard in the 1930s and in the most recent decade. The inference that climate conditions during that period were as warm as nowadays is indirectly corroborated by evidence stemming from the other types of proxy data from the Lomonosovfonna ice core. In particular, the degree of summer melt estimated from analysis of ionic washout indices (W_{NaMg}, W_{ClK}; see Grinsted et al. [2006] for details) suggests that warm summers prevailed before the 1300s. Repeated sampling at the drilling location on Lomonosovfonna during the 2000–07 field campaigns have demonstrated that such a degree of melt, as was observed in Medieval times, has been exceeded only in a very recent few years (Kekonen, Moore, Perämäki, Mulvaney et al. 2005; Virkkunen et al. 2007).

Conclusions

Calibration of the two isotopic ice-core series from western Svalbard has enabled us to quantify the temporal evolution of past winter temperatures in Svalbard. The approximately millennial-scale reconstruction of the Longyearbyen DJF SAT can be conditionally partitioned into three major sub-periods. The continuous winter temperature decline during the period of 800–1800, with a magnitude of about 0.9 °C century^{-1}, was followed by the coldest century in Svalbard according to the reconstruction. The scale of the Little Ice Age-associated winter cooling was on the order of 4 °C, as inferred from the comparison of the reconstructed and instrumental DJF temperatures for the two periods of the 1800s and 1900s. The same procedure carried out for Vardø instrumental temperature series yielded a negative trend of magnitude 0.3 °C century^{-1} before the 1800s and a Little Ice Age cooling of 1.3 °C on average in northern Norway. The rapid warming at the beginning of the 20th century is already well documented in the instrumental data and was accompanied by a parallel decline of sea-ice extent in the study area. However, both the reconstructed DJF SAT as well as indirect indicators of summer temperatures suggest that the Medieval period was at least as warm as the end of the 1990s in Svalbard.

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