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Deuterium excess record from a small Arctic ice cap

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[1] In this paper we present a deuterium excess (d) record from an ice core drilled on a small ice cap in Svalbard in 1997. The core site is located at Lomonosovfonna at 1255 m asl, and the analyzed time series spans the period 1400–1990 A.D. The record shows pronounced multidecadal to centennial-scale variations coherent with sea surface temperature changes registered in the subtropical to southern middle-latitude North Atlantic during the instrumental period. We interpret the negative trend in the deuterium excess during the 1400s and 1500s as an indication of cooling in the North Atlantic associated with the onset of the Little Ice Age. Consistently positive anomalies of d after 1900, peaking at about 1950, correspond with well-documented contemporary warming. Yet the maximum values of deuterium excess during 1900–1990 are not as high as in the early part of the record (pre-1550). This suggests that the sea surface temperatures during this earlier period of time in the North Atlantic to the south of approximately 45°N were at least comparable with those registered in the 20th century before the end of the 1980s. We examine the potential for a cold bias to exist in the deuterium excess record due to increased evaporation from the local colder sources of moisture having isotopically cold signature. It is argued that despite a recent oceanic warming, the contribution from this local moisture to the Lomonosovfonna precipitation budget is still insufficient to interfere with the isotopic signal from the primary moisture region in the midlatitude North Atlantic.

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1. Introduction

[2] Over the past few decades ice cores have become a powerful tool for studying the climate of the past. The most common method is the use of stable water isotopes $\delta^{18}O$ and δD as a proxy for the condensation temperature in the atmosphere at the time of precipitation. The respective variability of $\delta^{18}O$ and δD in the ice core can subsequently be translated into past surface air temperature changes. The approach has been successfully applied to ice core data from Antarctica [e.g., *Jouzel et al.*, 1987, 2007b; *EPICA Community Members*, 2004] as well as from Greenland [e.g., *Fischer et al.*, 1998; *Hoffmann et al.*, 2001; *North Greenland Ice Core Project Members*, 2004]. Although the physical fractionation processes are quite well understood [e.g., *Jouzel and Merlivat*, 1984; *Ciais and Jouzel*, 1994], the water stable isotopes in the ice core are integrated tracers of the water cycle. The actual $\delta^{18}O$ and δD in precipitation are affected by changes in evaporation conditions, atmospheric processes including convection along the air mass trajectory and condensation processes. Finally, the initial $\delta^{18}O$ and δD profiles in the ice core are subject to modifications due to various post deposition effects [*Fisher et al.*, 1985; *Koerner*, 1997]. The temporal and spatial robustness of the $\delta^{18}O$/T and δD/T relationships are also contested. On a glacial-interglacial time scale the bias due to changes in precipitation seasonality [*Cuffey et al.*, 1995; *Krinner and Werner*, 2003] as well as shifts in the water vapor source area [*Boyle*, 1997] have to be taken into account. Whereas on shorter, subannual time scales, the interpretation of the isotopic composition of the aggregate of individual accumulation events in terms of local temperature fluctuations can be arguable [*Helsen et al.*, 2006].

[3] The variability of the second-order parameter, deuterium excess $d = \delta D - 8\delta^{18}O$ defined by *Dansgaard* [1964], is largely governed by kinetic effect in the isotopic fractionation processes that take place out of thermodynamic equilibrium [*Jouzel and Merlivat*, 1984; *Ciais and Jouzel*, 1994]. Under a typical situation of undersaturated atmospheric water vapor (with respect to the oceanic surface),

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Figure 1. Map of the study area showing the drilling site at Lomonosovfonna and other locations referenced in the text.

the magnitude of this effect depends on sea surface conditions. A higher sea surface temperature (SST), lower relative humidity and, to a lesser extent, a higher wind speed, intensify evaporation. The kinetic effect is increased leading to a larger deuterium excess in the evaporated moisture. As a result, the deuterium excess data contain information on conditions prevailing in the oceanic source regions which provide moisture for polar precipitation. Variations in the moisture transport history [*Helsen et al.*, 2006] or/and kinetic fractionation occurring during snowflake formation [*Jouzel and Merlivat*, 1984] may exert an additional important local influence to the final value of d in precipitation.

[4] Deuterium excess records from ice cores from the interior of the large ice sheets of Antarctica and Greenland have thus been used as a link to the past ocean climate and interpreted in terms of changes in the moisture source temperature. This has been tested both on glacial-interglacial time scales [e.g., *Johnsen et al.*, 1989; *Stenni et al.*, 2001; *Vimeux et al.*, 1999, 2001] and throughout the Holocene [*Masson-Delmotte et al.*, 2005b]. Deuterium excess records from Greenland have furthermore been used to link the ice core data to more recent climate variability such as the cooling during Little Ice Age (LIA) [*Hoffmann et al.*, 2001] and large-scale atmospheric dynamics such as the North Atlantic Oscillation (NAO) [e.g., *Barlow et al.*, 1993]. Although records from polar glaciers outside the major ice sheets can be a valuable complementary source of information about past climate variations, only one such high-resolution deuterium excess record has been presented in the literature so far: the ice core record from Akademii Nauk Ice Cap (ANIC) on Severnaya Zemlya [*Fritzsche et al.*, 2005]. Recently it was demonstrated that despite strong summer melt, this record shows good correlation with the Kara Sea ice extent, pointing to this area as an important regional source of moisture for ANIC precipitation [*Opel et al.*, 2008]. In this paper we will present and discuss a deuterium excess record from a Svalbard ice core.

[5] Since the 1970s several ice cores have been drilled on the low-altitude glaciers and ice caps in Svalbard by groups from both the former Soviet Union [e.g., *Tarrusov*, 1992; *Kotlyakov et al.*, 2004] and Japan [*Watanabe et al.*, 2001], however few of these ice cores have been studied in great detail. In addition, interpretations are complicated by poor dating constraints, coarse sampling and limited chemical analyses, implying a limited climatological reconstruction of the area based on these older ice cores.

[6] During the last decade some new ice cores have been drilled on three of the ice caps on Svalbard; among them is one on the summit of Lomonosovfonna retrieved in 1997 [*Isaksson et al.*, 2001; *Kekonen et al.*, 2005] (see location in Figure 1). This ice core has been estimated to cover the past 800 years using a combination of dated reference layers, such as the 1963 ^{137}Cs peak and volcanic eruptions, annual cycles of water isotopes and glaciological modeling [*Pohjola et al.*, 2002a; *Kekonen et al.*, 2005].

[7] Seasonal melting of the snow pack represents a serious obstacle when extracting climate and environment-related records from ice cores situated below the dry firn zone [*Koerner*, 1997]. During warm events, percolation of water originating either from rain or from melting of the ice pack will tend to diffuse downward and advect any chemical or physical signal in the ice strata. Fractionation of the isotopic mix during refreezing of the percolated

water, reevaporation and diffusion of the water vapor in the firn all cause further alteration of the precipitated isotopic ratio. The warmer the site, the greater the destruction of the signal through these processes. The Lomonosovfonna ice core has therefore been thoroughly evaluated with respect to postdepositional alterations of the ice core records due to percolation of melt water during summer [*Pohjola et al.*, 2002b; *Moore et al.*, 2005]. These works suggest that during warm summers as much as 50% of the annual accumulation may melt and percolate into the underlying firn. For most summers this decreases to 25%. The water typically refreezes within one or, rarely, two annual layers [*Pohjola et al.*, 2002b]. However, even with melt percentages as high as 80%, *Moore et al.* [2005] show that there is little disturbance to the chemical stratigraphy of the Lomonosovfonna ice core. Percolation and diffusion in this ice core introduce uncertainties in ion locations resulting in errors approximately equal to the error in the dating model based on layer thinning between marker horizons. Most parameters preserve an annual, or in the worst cases, a biannual atmospheric signal.

[8] Stable water isotope records were shown to experience smaller postdepositional alterations due to summer melt compared with the leakage of ions in the snow and firn pack. This conclusion was reached by investigating the statistical properties of their concentration distributions for layers with different amounts of melting [*Pohjola et al.*, 2002b]. Furthermore, we observe cycles in most ice core parameters (see examples for $\delta^{18}O$, δD and d given by *Pohjola et al.* [2002b, Figure 8]), and the water isotope record has similar statistical characteristics as the isotopic records from Svalbard coastal stations [*Pohjola et al.*, 2002b]. A good agreement was found between Lomonosovfonna $\delta^{18}O$ and meteorological records from around the Barents Sea [*Isaksson et al.*, 2005; *Grinsted et al.*, 2006] as well as the Lomonosovfonna borehole temperature profile [*van de Wal et al.*, 2002]. Hence we conclude that the Lomonosovfonna ice core, despite melt events and percolation, contains a record that can successfully be used in climate and environmental studies [*O'Dwyer et al.*, 2000; *Isaksson et al.*, 2005; *Grinsted et al.*, 2006; *Moore et al.*, 2006; *Virkkunen et al.*, 2007].

[9] The question we will consider in this paper is "What climatic information can be extracted from the deuterium excess record from the Lomonosovfonna ice core?" We test this time series against available instrumental sea surface temperature and sea level pressure (SLP) data, as well as modern monitoring of the isotopic composition of precipitation on Svalbard, and then use it to infer past climate variability. The fact that Svalbard is situated in an area of large thermal gradients associated with the vicinity of the Arctic front, makes the ice core records from this area valuable as potential indicators of large-scale climate changes.

2. Data Series and Analysis

[10] The 121 m long ice core was drilled using an electromechanical drill (diameter of 105 mm) on Lomonosovfonna (Figure 1). This is one of the highest ice fields in Svalbard located at 1255 masl. The current estimated annual temperature range is from 2°C to about −40°C; the in situ temperature measurements are however very scarce. Total ice depth from radar sounding was 123 m, and the site is close to the highest point of the ice cap with roughly radial ice flow. The mean annual accumulation rate was estimated to be around 0.30 m w.e. for the period 1715–1950, with an increase to an average of 0.41 m w.e. for the period 1950–1997 [*Pohjola et al.*, 2002a].

[11] A total of 2300 oxygen isotope samples were analyzed. The ice core was cut into 5 cm sections using a band saw in the cold laboratory. The details are described by *Isaksson et al.* [2001] and a thorough discussion of the water isotopes in relation to melt features is presented by *Pohjola et al.* [2002b]. For the $\delta^{18}O$ we used standard procedures for oxygen isotope preparations [*Epstein and Mayeda*, 1953]. Conversion of H_2O to CO_2 and its subsequent extraction and analysis is done on a fully automated system [*Stahrenberg*, 1985] attached to a Finnigan-MAT Delta-E mass spectrometer. For about 10% of the samples a replicate analysis was carried out with an overall reproducibility better than ±0.1‰.

[12] The deuterium analysis is based on the method originally described by *Bigeleisen et al.* [1952]. We selected a subset of 1230 samples from those previously analyzed for the oxygen isotope content. A sample of 10 μl of water is quantitatively reduced to pure hydrogen gas on a hot uranium surface. To reduce memory effects, each sample is injected and reduced twice, and only the second aliquot is used. The hydrogen gas is analyzed for its deuterium content using a SIRA 9 dual inlet Isotope Ratio Mass Spectrometer. Corrections for H_3^+, and for cross contamination [*Meijer et al.*, 2000] have been applied. The deuterium scale of the mass spectrometer is calibrated with respect to VSMOW and normalized using SLAP [*Coplen*, 1988], both on a frequent basis. The overall accuracy for δD measurements is estimated to be ±1.5‰. This yields the root mean square instrumental error for an individual estimate of d of about Δd = ±1.7‰

[13] Figure 2 shows the time series of the oxygen and hydrogen isotopes used to calculate the deuterium excess (the raw time series are plotted in Figure 2, bottom). The sampling resolution generally increases with time: from roughly annual or biannual prior to 1815, to subannual, with 3–7 samples per year on average, after 1815 (Figure 3a). The dating uncertainty is about 20 years as inferred from the comparison of modeling and layer counting for the year 1700 [*Kekonen et al.*, 2005]. The quantitative analysis we present here is restricted, however, to the most recent period of 150 years, where the dating error is estimated to be only 3 years. The raw time series were tested for the presence of outliers using the Quartile (or Fourth-Spread) method. There are 5 out of 1056 samples (1 out of 740 for the period 1850–1997) in total that fall outside the admissible range of [−4.5, 22.9]‰ calculated from [Q_1 − 3 * IQR, Q_3 + 3 * IQR]. Here IQR denotes the interquartile range equal to the difference between the third (Q_3) and first (Q_1) quartiles of the analyzed data. The outliers appear as sporadic spikes in the record and were eliminated from the analysis.

[14] For further analysis we resampled the raw time series by aggregating and averaging all d values from samples ascribed to the same calendar year according to the established depth-age chronology. We then subtracted the mean

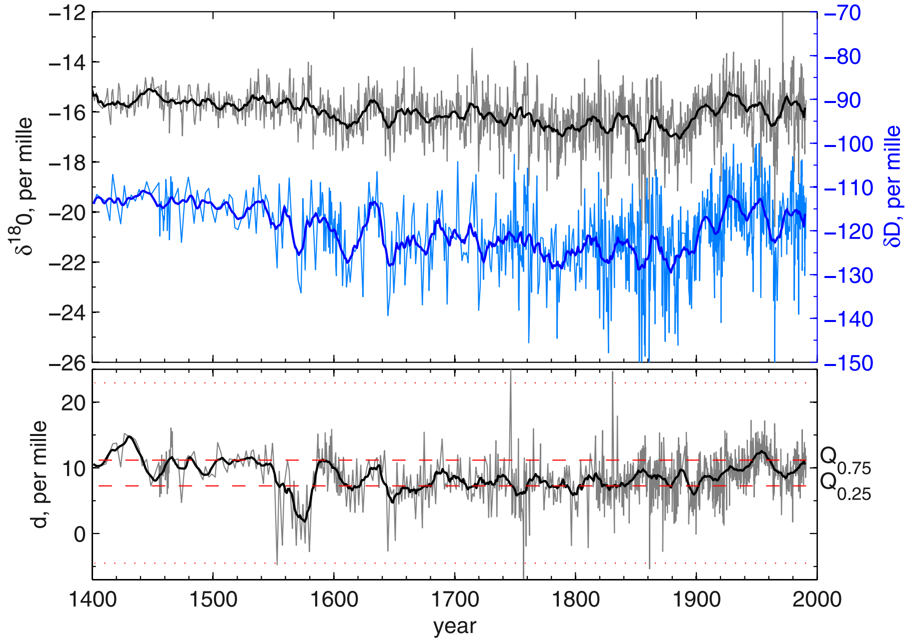

Figure 2. Isotope records of the Lomonosovfonna ice core. (top) Raw Lomonosovfonna δ^{18}O (grey) and δD (light blue) time series together with their 11-year running means estimated from annually resampled data (left and right axes, respectively). (bottom) Raw Lomonosovfonna deuterium excess and its 11-year running means. Dashed red lines mark the first ($Q_{0.25}$) and the third ($Q_{0.75}$) quartiles of the data series; data points above and below the dotted red lines were considered to be outliers.

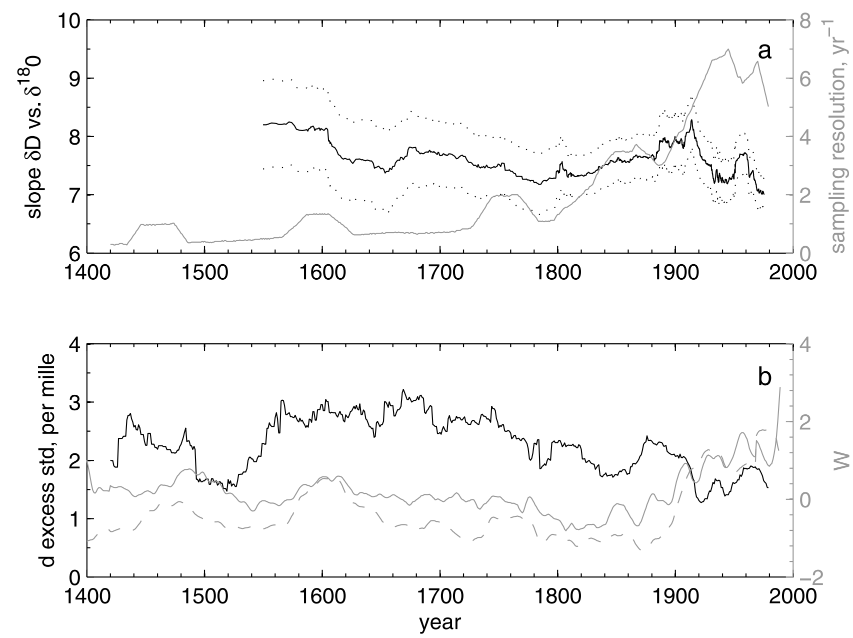

Figure 3. (a) Temporal variations of the slope of the Lomonosovfonna local meteoric water line, as estimated by sliding a window of 200 raw values over the *dD* and $\delta^{18}O$ series (black solid line, left axis). Dotted lines cover the interval within ±STD of the slope. The actual width of the sliding intervals in the time domain is inversely proportional to the sampling resolution of the deuterium excess series (grey line, right axis). It scales down from approximately 300, in the early part, to 30 years in the top of the core. (b) Temporal variability of the standard deviation of the deuterium excess (left axis, black line) and dimensionless ionic washout indices recording summer melt at the summit of Lomonosovfonna (right axis, solid grey line is W_{NaMg}, and dashed grey line is W_{ClK}, see *Grinsted et al.* [2006] for definition and further details). All values are estimated in 40-year sliding intervals.

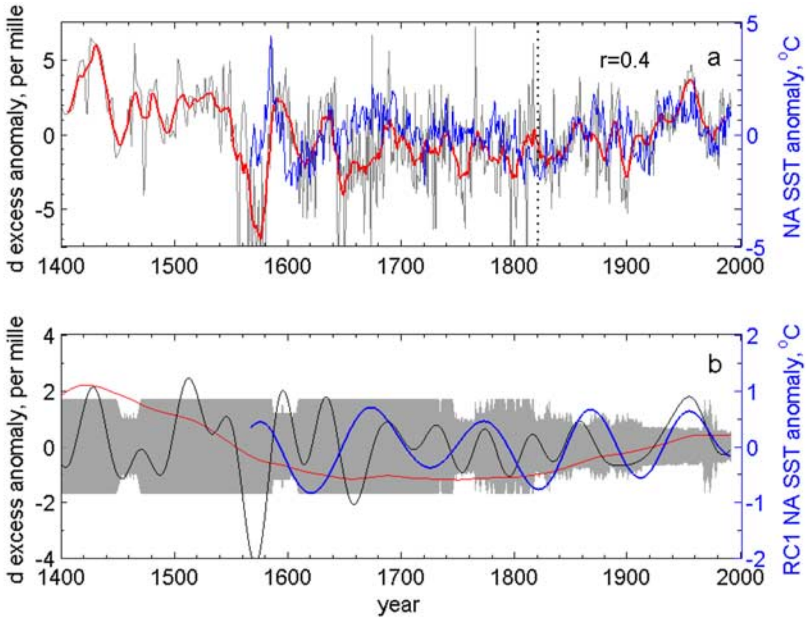

Figure 4. (a) Annual mean deuterium excess anomalies for Lomonosovfonna together with the 11-year running average (red line). Solid blue line in Figure 4a shows the reconstructed mean annual SST anomalies in the North Atlantic. r designates the correlation coefficient between the reconstructed SST and Lomonosovfonna d for the period after 1820, as highlighted by the vertical black dotted line. (b) SSA analysis (window length = 120) of Lomonosovfonna d. Red and black lines highlight the reconstructed nonlinear secular trend and centennial (RC1) plus 40-year (RC2) variability, respectively. Solid blue line from the right shows the SSA-derived (window length = 120) multidecadal RC1 of the reconstructed North Atlantic SST anomalies. Grey shading in the background shows the accuracy of the estimated annual mean deuterium excess.

(d_{mean} = 8.9) to produce mean annual anomalies and filled the remaining gaps in the data by spline interpolation. There are 162 interpolated data points in total, mainly prior to 1800, where the annual resolution is not always attained, and only 4 interpolated points in the period 1854–1990. The resulting time series of anomalies is shown in Figure 4. To highlight the decadal variations in deuterium excess we smoothed the series using an 11-year running mean. The accuracy of the annual mean estimates is calculated as $\Delta d/\sqrt{N_y}$, where N_y denotes a number of samples for a particular year (shown in Figure 4b).

[15] Annual mean dD and $\delta^{18}O$ for Lomonosovfonna are highly correlated with a r^2 value of 0.82 (0.86 for the raw data). The deuterium excess, however, carries an additional information compared with the dD and $\delta^{18}O$ series. This is supported by the principal component analysis performed on the two isotopic time series, following the method applied by *Vimeux et al.* [1999] for analysis of the Vostok ice core data. The first principal component, PC1, is reminiscent of the variations in dD and $\delta^{18}O$, while PC2 which is orthogonal to PC1 is coherent with the deuterium excess (r^2 = 0.98). Despite the fact that neither the raw nor the annual mean series of $\delta^{18}O(dD)$ are correlated with d ($r{\approx}0.04$), they do share common features on longer scales. Figure 5 shows 50-year running means of d plotted as a function of the smoothed $\delta^{18}O$. Figure 5 reveals an evident tendency for the pairs of (d, $\delta^{18}O$) to cluster in two distinct groups. In the earlier (before approximately 1550) and the most recent (after 1990) parts of the record a higher deuterium excess concurs with a higher $\delta^{18}O$. Lower d and $\delta^{18}O$ values are correspondingly observed between 1550 and 1900. Hence on centennial scales d and $\delta^{18}O$ in the Lomonosovfonna core are positively correlated. It is worthwhile to note that, on much longer, glacial-interglacial

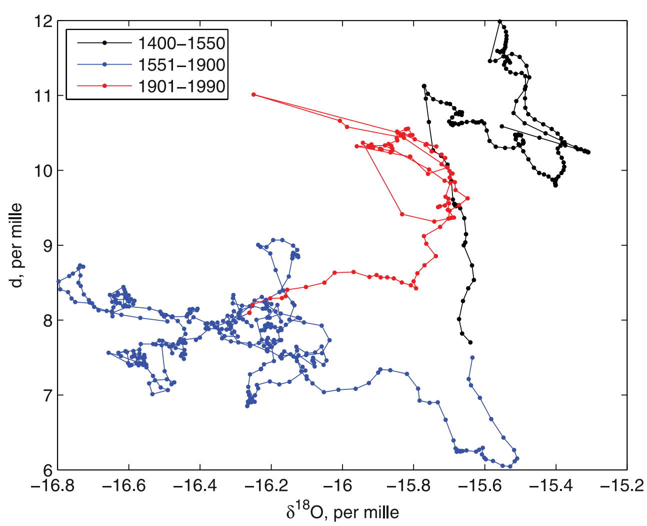

Figure 5. Lomonosovfonna deuterium excess versus δ^{18}O; the records are smoothed using 50-year running means. The color highlights three different periods with distinct values of d/δ^{18}O relationship: before 1550 (black), 1550–1900 (blue), and after 1900 (red).

Table 1. Statistical Characteristics of Lomonosovfonna Deuterium Excess Record[a]

Period (years)	n (n_{raw})	μ	σ	$Q_{0.25}$	$Q_{0.75}$	ζ	κ
1400–1990	422 (1173)	8.9	2.6	7.0	10.5	0.01	−0.15
1400–1550	60 (80)	11.0	2.2	9.5	12.5	−0.11	0.05
1551–1882	257 (480)	8.1	2.5	6.5	9.7	0.27	0.25
1883–1990	105 (614)	9.5	2.0	8.2	10.8	−0.48	0.03
1840–1970	129 (619)	9.2	2.1	7.6	10.6	−0.17	−0.43

[a]n_{raw}(in parentheses) denotes the number of raw samples available for the reference period, and n is the number of derived annual means used for calculating the time series parameters; μ and σ denote the mean and standard deviation, respectively; ζ is the skewness of each distribution, with high absolute values for a skewed series and low values for less skewed series; κ is kurtosis, which has a value of 0 for perfectly distributed Gaussian data but gives higher values for a more spiky distribution; and $Q_{0.25}$ and $Q_{0.75}$ are the average of the first quartile (the 25% highest numbers) and of the fourth quartile (the 25% lowest numbers), respectively, of the distribution.

scales, this relationship was found to have the opposite sign for ice core data from Antarctic [*Vimeux et al.*, 1999].

[16] The local meteoric water line (LMWL) for the annual means was calculated to be $dD = 8.03(0.18)d^{18}O - 9.06(2.96)$ for the entire period under consideration, 1400–1990, and $dD = 7.82(0.09)d^{18}O - 6.15(1.48)$ for the raw data. The numbers in parentheses indicate the standard deviations (STD) of the slope and y intercept estimates. The calculated parameters of the LMWL are very close to the globally averaged precipitation line (global meteoric water line, GMWL), which has a slope of 8 [*Dansgaard*, 1964] and the earlier estimate of 7.5 made by *Pohjola et al.* [2002b] for the two shorter time periods of 1910–1937 and 1982–1993. In order to check if the slope changes over time, we estimated its magnitude by sliding a window of 200 raw values over the dD and $\delta^{18}O$ series. The results presented in Figure 3a suggest the slope is relatively stable, experiencing only subtle variations in time and being confined to the interval of [7,8.2]. It reaches a local maximum of about 8.2 after 1900, followed by a gradual decrease to the present day value close to 7.

[17] Even though the isotopic profiles from the analyzed core seemed to experience relatively moderate melt-, percolation- and evaporation-related alterations, these processes might exert an amplified effect on the noise in the constructed deuterium excess series. A quantitative assessment of the role of the kinetic effect under phase transitions in the ice matrix is complicated because of the complexity of the associated processes. Their integral effect can however be evaluated by comparing the distribution of d for different classes of ice facies. Analysis presented by *Pohjola et al.* [2002b] revealed no substantial (i.e., above the instrumental error) discrepancy in d between the segments of the core which experienced little or no infiltration of meltwater and those heavily affected by melt. In addition, parameters of the LMWL are relatively stable in time and close to the GMWL, indicating that the role of kinetic fractionation within the ice is not substantial enough to alter the deposited d profile. Averaging the Lomonosovfonna ice core data over 5-year time intervals was shown to be sufficient to minimize the effects of melting, infiltration and refreezing on isotopic profiles [*Grinsted et al.*, 2006]. Our results suggest however, that even the use of the nonsmoothed time series still provides meaningful results.

[18] To make inferences on which processes drive the variability of deuterium excess and to examine possible teleconnection patterns, we use Hadley Centre's mean SLP data (HadSLP2 [*Allan and Ansell*, 2006]) and Extended Reconstructed SST data from NOAA NCDC ERSSTv2 [*Smith and Reynolds*, 2004]. The data are available in the form of monthly means and span the periods 1850–2004 (SLP) and 1854–2004 (SST). The horizontal resolutions are 5° latitude by 5° longitude and 2° by 2°, respectively.

3. Results and Discussion

3.1. Lomonosovfonna Deuterium Excess Record: Overall Characteristics

[19] Basic statistical characteristics of the Lomonosovfonna d series are presented in Table 1. The data are shown for the whole record and four selected subperiods: 1440–1550 (d, δD and $\delta^{18}O$ are consistently above the average for the whole record), 1551–1882 (d, δD and $\delta^{18}O$ are below the average), 1883–1990 and 1840–1970. In the latter two cases there are deuterium excess data available for locations in the European Arctic.

[20] The mean d value of 9.2‰ estimated for a period of 1840–1970 is lower than the 11.0‰ reported for high-resolution GISP2 deuterium excess time series from Greenland [*Barlow et al.*, 1993]. This is also less, 9.5‰ versus 11.0‰, than found for the low-altitude (800 m asl) ANIC ice core for the period 1883–1990 [*Fritzsche et al.*, 2005]. This points to a more "marine" location of Lomonosovfonna implying shorter distillation paths for moisture arriving at the core site [*Pohjola et al.*, 2002b]. We presume that the higher average d of the ANIC core, despite its lower elevation, is due to the greater remoteness of Severnaya Zemlya from major moisture sources, longer sea ice season and hence, longer distillation paths for the moisture forming precipitation at the ANIC core site.

[21] The comparative analysis of Lomonsovfonna d with isotope records from precipitation sampled at coastal locations on Spitsbergen is hampered by the limited period of the observational series. The available GNIP records from Svalbard stations Isfjord Radio and Ny-Ålesund span the periods of 1961–1976 and 1990–2002, respectively (International Atomic Energy Agency/World Meteorological Organization, Global Network of Isotopes in Precipitation database, available at http://isohis.iaea.org). One should note that in the first case, the observations are rather irregular and the annual means can only be robustly estimated for 7 years. The Ny-Ålesund data have only 1 year overlap with the Lomonosovfonna time series. Still, the comparison of annual mean deuterium excess from the ice core with the direct measurements suggests somewhat higher deuterium excess values in Lomonosovfonna (9.6‰ versus 8.9‰ for 8 years of overlap), which is likely

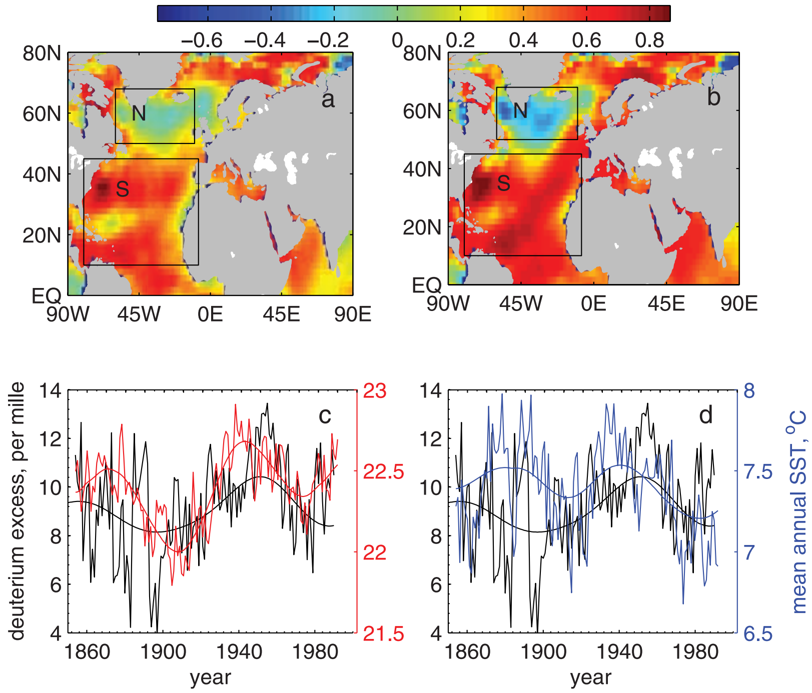

Figure 6. (a and b) Correlation maps between gridded mean annual North Atlantic sea surface temperature and Lomonosovfonna deuterium excess. The data series were smoothed using 3 (Figure 6a) and 21 (Figure 6b) year running means. Color bar at the top shows the magnitude of the correlation. Correlation coefficients with a magnitude above 0.29 (Figure 6a) and 0.71 (Figure 6b) are statistically significant at the 95% level. (c and d) Lomonosovfonna deuterium excess (black line) and box S (red line in Figure 6c) and N (blue line in Figure 6d) averaged annual mean SST. Smooth solid lines of the respective colors are the SSA-derived nonlinear trend components.

due to the higher-altitude location of the core site [*Pohjola et al.*, 2002b].

[22] The overall magnitude of interannual variability in Lomonosovfonna deuterium excess is well characterized by its standard deviation. In order to check how it varies in time we calculated STD_d in 40 years sliding intervals (Figure 3b). Figure 3b shows a gradual decrease in variability over the course of the last centuries, occurring in parallel with a warming during this period, as suggested by isotope records and various historical evidence [*Isaksson et al.*, 2005; *Grinsted et al.*, 2006; *Bradley et al.*, 2003]. The result is in agreement with the conclusion drawn by *Masson-Delmotte et al.* [2005b] suggesting a less stable climate associated with generally colder conditions.

3.2. Association With Sea Surface Temperatures and Sea Level Pressure During the Instrumental Period

[23] We performed a correlation analysis of deuterium excess time series with reconstructed annual mean SST data during the common period of 1854–1990. In order to minimize the effect of possible dating errors, a 3-year running mean was applied to the deuterium excess time series and the SST data prior to analysis. Figure 6a reveals a complex pattern with the largest correlation coefficients reaching 0.6 in the North Atlantic to the south of 45°N. The maximum correlation is reached to the east of the North American coast between 30°–40°N which is known to be an east coast cyclogenetic area [*Whittaker and Horn*, 1981]. Noteworthy is the area with no correlation to the south of Greenland and Iceland. This is also a region of frequent cyclogenesis [*Serreze et al.*, 1997] and is known to sustain an anomalous heat and mass exchange at the atmosphere-ocean interface associated with the sinking branch of the ocean's overturning circulation [*Marshall et al.*, 2001].

[24] When smoothing the data using a 21-year running mean, the positive correlation of d with the North Atlantic SST to the south of 45°N further increases (see Figure 6b). The region with negative correlation to the south of Greenland and near Iceland is now clearly evident, although its magnitude is lower compared with the positive correlation maximum.

[25] The area-weighted annual mean SST in the box 80°–7.5°E, 10°–45°N (denoted S) shows reasonably good agreement with Lomonosovfonna d, as characterized by a correlation coefficient of 0.44 (statistically significant with the 95% confidence). These results support a concept of the deuterium excess capturing the SST variability at the remote source of moisture forming precipitation on Lomonosovfonna.

[26] Little is known about the composition of precipitation on Svalbard ice caps. Since the archipelago is located on pathways of Arctic low-pressure systems [*Whittaker and Horn*, 1984; *Serreze et al.*, 1997], it is reasonable to suppose that most of the precipitation is associated with cyclonic activity. Although the majority of cyclones reaching Svalbard have their origin in the Icelandic Low off the southeast coast of Greenland [*Whittaker and Horn*, 1984; *Serreze et al.*, 1997], they carry vapor that may have originated as far

south as 30°N. In Figure 6 the highest correlations are found from subtropical to southern parts of the midlatitude North Atlantic. This is also the region identified in modeling and observational studies [*Werner et al.*, 2001; *Johnsen et al.*, 1989] as the chief supplier of moisture to the polar and subpolar North Atlantic. The moisture evaporated in this area is efficiently entrained by cyclonic systems forming along the east coast of North America, which then transport it northward in the so-called warm conveyor belts. One should note that the entrainment of the isotopically cold high-latitude moisture in the cyclogenetic area to the south of the coast of Greenland is not ruled out. With "isotopically cold" we refer to the moisture evaporated from a colder sea surface and hence characterized by a decreased value of deuterium excess, in contrast to "isotopically warm" which is evaporated from a warmer surface. Its relative contribution to the total moisture budget, however, varies from year to year, as indicated by zero correlation (negative on longer time scales) between SST and *d*, particularly for this area.

[27] We use singular spectrum analysis (SSA [*Ghil et al.*, 2002], window length = 30) to extract nonlinear trends from the data over the common instrumental period (Figures 6c and 6d). The SSA-derived multidecadal variation in Lomonosovfonna *d* and the North Atlantic SST in box S, over the analyzed 140-year period, have most power at different time scales. The SST_S exhibits power at a shorter period (about 70 years), whereas the deuterium excess tends to show a significant power at about a 90 year time scale. Also, the minimum in SST_S at around 1910 lags the minimum in deuterium excess by approximately 20 years. Note that *d* during 1880–1900 exhibits stronger variability compared with the rest of the considered period, as seen in Figure 3b. The observed increase of *d* since 1920 matches an increase in SST_S, but is interrupted by a decrease in the 1930s. The resulting maximum in the deuterium excess at about 1950 is less broad and lags by 10 years the 1940 maximum in SST_S. Analysis of area-weighted annual SST in the box 60°–10°E, 50°–68°N (denoted N) in Figure 6d reveals pronounced positive SST_N anomalies during the two periods: 1870–1900 and 1930s (see Figure 6d) implying a negative correlation between SST_N and *d*, in agreement with Figure 6b.

[28] Cooling (warming) in the northern middle latitudes of the North Atlantic (north of approximately 45°N), efficiently decreases (increases) evaporation there. We suggest that it introduces a bias in the deuterium excess signal in the Lomonosovfonna precipitation, changing the relative fractions of moisture originating from the warmer (sub)-tropics and southern midlatitudes to the south of 45°N and colder northern midlatitudes. Lower values of *d* during 1880–1900 and the 1930s (and hence weak negative correlations of SST and Lomonosovfonna *d* on longer time scales in the region to the south of Greenland) may therefore be indicative of an effect of a smaller fraction of isotopically warm moisture in the atmospheric moisture composition and precipitation budget on Lomonosovfonna during these periods.

[29] A pronounced retreat of sea ice in the Nordic Seas since approximately 1850 [*Vinje*, 2001; *Divine and Dick*, 2006] and the overall warming of the North Atlantic [*Polyakov et al.*, 2005] shows the potential for the overall shift in the moisture composition in Svalbard precipitation to a higher fraction of isotopically "cold" moisture. This mechanism exerts a substantial influence on deuterium excess variations, as was recorded in ice cores from Greenland and Antarctica on glacial-interglacial time scales [*Boyle*, 1997; *Vimeux et al.*, 1999; *Jouzel et al.*, 2007a; *Masson-Delmotte et al.*, 2005a] and in the Holocene [*Masson-Delmotte et al.*, 2005b; *Hoffmann et al.*, 2001]. The estimated trends in the normalized annual mean northern North Atlantic (30°N–70°N) SST and deuterium excess for the period 1900–1990 are however close within the error, $0.013 \pm 0.004\ a^{-1}$ and $0.011 \pm 0.004\ a^{-1}$, respectively. This leads us to conclude that, at least on the time scale of the last century, the contribution of this effect is not large enough to be captured in the record.

[30] An analysis of composites of SLP and SST fields for 1854–1990 based on selected cases of anomalously high and low values of *d* further confirms the results derived from the correlation analysis. We constructed composites of winter (DJF) SST and SLP anomalies for years in which the magnitude of the deuterium excess anomaly exceeded $\pm 1.0\ d_{STD}$. Figure 7a shows the difference between the $+d$ and $-d$ composites. The use of winter months for analysis is justified by more prominent spatial patterns of SST and SLP compared with other seasons. Possible causes include higher variability in the amount of winter precipitation, as was found for Svalbard airport data [*Hanssen-Bauer and Førland*, 1998], seasonal (winter) peak in precipitation at higher altitudes [*Rasmussen and Kohler*, 2007] and/or higher deuterium excess variability in winter precipitation as recorded in Ny-Ålesund GNIP data. The latter agrees well with generally higher winter SST variability in the potential moisture source region, the North Atlantic. In the North Atlantic the analysis reveals the monopolar pattern in SST with the maximum temperature difference in the zone between 30° and 50°N. Its configuration generally resembles that of the warm phase of the Atlantic Multidecadal Oscillation (AMO) [*Knight et al.*, 2005; *Grosfeld et al.*, 2007], a coherent pattern of multidecadal variability in sea surface temperature centered on the North Atlantic Ocean, but with lesser northward extent. The latter is expected as the SST just to the south of Greenland shows no correlation with Lomonosovfonna *d* on interannual time scales (see Figure 6a). The structure of the respective SLP pattern in Figure 7a also has an AMO-like signature but with a transition from positive to negative anomalies shifted somewhat northward compared with the pattern shown by *Grosfeld et al.* [2007].

[31] We applied the same composite analysis to the isotope record from precipitation sampled at Ny-Ålesund during 1990–2002. The inferred signature of the observed positive/negative *d* anomalies in SST and SLP fields in the winter is shown in Figure 7b. The analysis reveals principally the same spatial structure of winter SLP and SST anomalies as found for the ice core *d* series in Figure 7a. It is noteworthy however, that the center of the maximum difference in SST in the North Atlantic is shifted slightly northward in Figure 7b compared with Figure 7a, suggesting that the moisture source for Svalbard precipitation might also have shifted northward during this period of time. Overall these results point to the similarity of the deuterium excess in precipitation and from the ice core data and

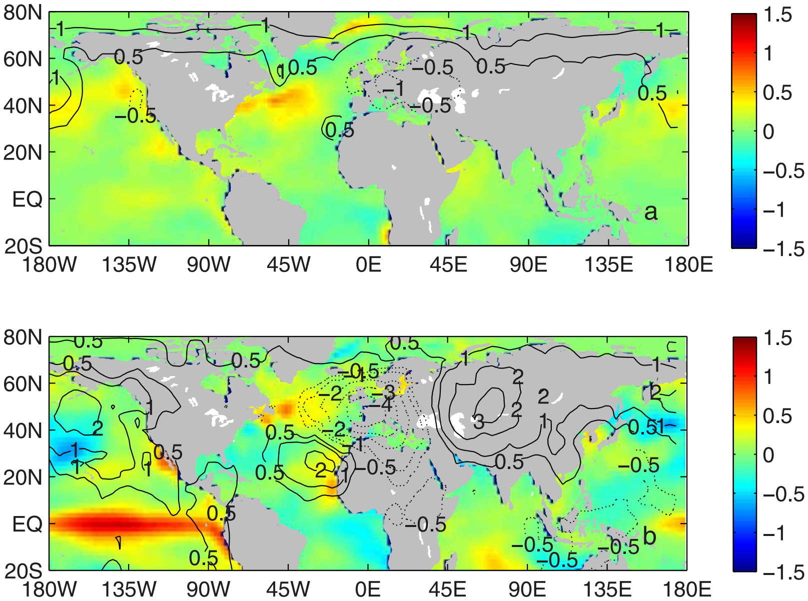

Figure 7. (a) Differences between composites in DJF SST (shown in colors) and SLP anomalies (contours) for periods with positive (larger than 1.0 STD) and negative (less than 1.0 STD) Lomonosovfonna *d* anomalies during 1855–1990. (b) Same as in Figure 7a but for the measured deuterium excess in DJF precipitation in Ny-Ålesund in years 1992, 1993, 1997, 1998, 2001, and 2002 (positive anomalies) and 1990, 1994, 1995, 1996, 1999, and 2000 (negative anomalies).

suggests that, despite the difference in altitudes between the two sites, the variability of deuterium excess is governed by similar physical mechanisms.

3.3. Long-Term Variations and Multidecadal Cycles in Lomonosovfonna Deuterium Excess

[32] A simple linear fit to the Lomonosovfonna deuterium excess series does not reveal any significant trend on the time scale of the whole record. Still, one can distinguish a tendency toward consistently positive anomalies in the earlier part of the record, (pre-1550) and again after 1920 (see Figure 4 and Table 1) with a prominent peak around 1950. Given the association with sea surface temperatures demonstrated in section 3.2, this implies a warmer moisture source during these periods. The positive anomaly of *d* in the 1950s is the longest and strongest during the period 1900–1990, which is consistent with the temporal evolution of the North Atlantic SST [*Polyakov et al.*, 2005]. Yet the most pronounced decadal-scale features in the whole record are the prolonged maximum of *d* centered on 1430 and the minimum in the 1560s. These suggest that even warmer/colder conditions that might have prevailed in the North Atlantic during these decades than in the well-documented SST maximum in the 1950s or minimum in the 1900s. This conclusion, however, must be viewed with caution because of less certain dating in the earlier part of the record and a potential for the cold bias associated with the moisture originating from the area to the south of Greenland.

[33] In order to quantify significant long-term changes and identify possible regular variations in Lomonosovfonna *d*, we performed singular spectrum analysis (window length = 120) on the record. The Monte Carlo significance test against the red noise background, having the same mean, variance and first-order autoregressive parameter as the *d* time series, revealed one nonlinear trend and two reconstructed components (RC) statistically significant at the 95% level. The results of the analysis are displayed in Figure 4.

[34] The nonlinear trend component represents some 17% of the variability and has a magnitude of about 3‰. The curve shows a gradual transition from a positive deuterium excess anomaly, in the earlier part of the record, to consistently negative values during the 1600s, 1700s and 1800s, followed again by positive anomalies in the 20th century. The onset and timing of this cooling is in good agreement with those inferred from the oxygen isotope data from the northern Greenland ice cores [*Fischer et al.*, 1998] and ice cores from Eurasian Arctic glaciers [*Kotlyakov et al.*, 2004]. It also fits well with the present understanding of the climate variability of the past millennium [*Bradley et al.*, 2003; *Juckes et al.*, 2006; *Jansen et al.*, 2007] with a generally colder period associated with the Little Ice Age followed by a subsequent recovery that commenced after 1850.

[35] The 19th century seemed to be the coldest period on Svalbard, as evidenced by $\delta^{18}O$ and δD series [*Isaksson et al.*, 2005], increased sea ice extent [*Vinje*, 2001; *Divine and Dick*, 2006] and continentality [*Grinsted et al.*, 2006] of the area, decreased summer melt [*Pohjola et al.*, 2002b] and ion mobility [*Grinsted et al.*, 2006] (Figure 3b). However, the deuterium excess record shows a broader LIA-associated minimum which indicates a colder North Atlantic throughout most of the 17th and the 18th centuries, in agreement with western European evidence for the coldest part of the Little Ice Age. This lag of some 5 decades between the onset of cooling on Svalbard and in the midlatitude North Atlantic can be clearly seen in Figure 5 as a period of relatively stable higher values of $\delta^{18}O$ and progressively decreasing *d*. It corroborates the conclusion of the Lomo-

nosovfonna deuterium excess record being more representative of the remote source region temperature evolution rather than of local Arctic conditions.

[36] A stacked high-resolution deuterium excess record from the Greenland cores GRIP and S93 [*Hoffmann et al.*, 2001], does not agree well with the Lomonosovfonna *d* series. Both records suggest colder moisture sources in the 1600s compared with the 1500s, but demonstrate the opposite trends afterward. Persistently positive *d* anomalies during 1750–1850 in GRIP are mirrored by the negative ones in the Lomonosovfonna *d*. This is followed by a positive trend in *d* in the Lomonosovfonna deuterium excess since approximately 1850, but a negative one in the GRIP series. A similar negative trend detected in the deuterium excess record from the neighboring GISP2 core [*Barlow*, 1994] suggests that this feature appears to be a regional signal. The reason for such dissimilarity remains unclear, but can be related to a relatively higher contribution of cold local sources of moisture in the Greenland precipitation budget.

[37] The analyzed time series exhibits pronounced multidecadal variations at the scales of about 100 (denoted RC1) and 40 (RC2) years. These two quasi cyclic components, also displayed in Figure 4b, account for nearly 11% and 9% of the total variance with a total magnitude of some 2.5‰. A simple scaling of the RC1+RC2 components in *d* to the long-term variations in the midlatitude North Atlantic (Figure 6) yields associated annual mean SST anomaly variations of the order of 0.4°C. Analysis of the nonlinear trend component in the deuterium excess series suggests that cooling of approximately the same scale could have taken place in the North Atlantic during the Little Ice Age.

[38] The good agreement found between the SST in the North Atlantic and Lomonosovfonna *d* during the instrumental period (section 3.2), implies that a similar correlation with reconstructed North Atlantic SST is anticipated. We use the tree-ring-based reconstruction of SST anomalies in the North Atlantic [*Gray et al.*, 2004]. During the calibration period of 1856–1990 it represents some 40% of the variability in the mean annual SST in the North Atlantic south of 70°N and covers the period of 1567–1990. Figure 4a suggests that the Lomonosovfonna *d* and reconstructed annual mean North Atlantic SST are positively correlated after approximately 1820, as indicated by the correlation coefficient of $r^2 = 0.16$ for the annual means ($r^2 = 0.35$ for 3-year running means). The agreement between the two series is largely due to common multidecadal variability. In order to isolate statistically significant periodic components in the North Atlantic SST we applied SSA with the same input parameters as *d*. The resulting RC1 component, shown in Figure 4b, has the period of approximately 90 years and represents some 22% of the total variability in the reconstructed North Atlantic SST. The two reconstructed multidecadal components are nearly in phase during 1760–1990, but diverge substantially in the preceding period.

[39] The relationship between the raw series in the preinstrumental part of the records is correspondingly less certain. Still, the two pronounced positive anomalies in the two records around 1590 and 1640 appear to be similar in both records. Both series also seem to share similar long-term features (common periods of increase and decrease) during 1640–1720. The latter results need, however, to be interpreted with caution as the time series considered represent two fundamentally different types of proxies, each with its own set of shortcomings. In particular, the dating of the Lomonosovfonna core is less certain below the Laki horizon of 1783, which is also the period when the multidecadal components begin to diverge. The tree-ring-based reconstructions, despite having accurate dating, suffer from the so-called "segment length curse" [*Cook et al.*, 1995] that affects the quality of the reconstruction by suppressing the low-frequency components.

4. Conclusions

[40] We have presented a 590-year-long deuterium excess record from the Lomonosovfonna ice field on Svalbard and explored its potential for use in paleoclimate studies. Our analysis shows a close similarity between the temporal variations of SST in the subtropical to southern midlatitude North Atlantic and the deuterium excess anomalies in the Lomonosovfonna ice core for the instrumental period. It suggests that this part of the North Atlantic plays a major role in the formation of the moisture budget for precipitation in Svalbard, in agreement with the isotopic distillation modeling-based estimates of *Johnsen et al.* [1989].

[41] The Lomonosovfonna deuterium excess record, which we therefore consider to be a proxy for the subtropical to southern midlatitude North Atlantic SST, reveals a gradual cooling during the 1400s and 1500s, associated with the onset of the Little Ice Age. A distinct cool period, as inferred from persistently negative anomalies of *d*, lasted for some 3 centuries. A subsequent warming commenced at the end of the 19th century and the maximum values of *d* were reached in the 1950s, in parallel with well-documented oceanic warming for that period [*Polyakov et al.*, 2005]. The proposed interpretation of the Lomonosovfonna deuterium excess series is therefore much in line with the current understanding of climate variability in the North Atlantic and Europe during the past millennium [*Bradley et al.*, 2003; *Juckes et al.*, 2006; *Jansen et al.*, 2007]. We note that the values of deuterium excess before the 1550s are higher than the maximal ones observed during the 20th century. It suggests that the sea surface temperatures in the North Atlantic to the south of 45°N were at least comparable with those registered in the 20th century before the end of the 1980s.

[42] The inferred centennial-scale variations in *d* are somewhat longer than found in the instrumental SST data. We attributed this to the effect of a higher contribution of isotopically cold moisture during the periods of warming in the area to the south of Greenland. Overall warming of the ocean over the past century promotes increased evaporation in the (sub)polar areas. This may potentially influence the isotopic composition of the moisture forming Svalbard precipitation, increasing the role of cold local moisture sources. The comparison of the recent trends in the records of deuterium excess and SST suggest, however, that the associated cold bias in the deuterium excess record should be negligible.

[43] We conclude thereby that, despite all the complications associated with postdepositional alterations in the snow pack at relatively low altitudes, our results further

confirm the potential of ice cores from small ice caps for analysis and reconstruction of the climate history.

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References

Allan, R., and T. Ansell (2006), A new globally complete monthly historical mean sea level pressure data set (HadSLP2): 1850–2004, *J. Clim.*, *19*, 5816–5842.

Barlow, L. (1994), Evaluation of seasonal to decadal scale deuterium and deuterium excess signals, GISP2 ice core, Summit, Greenland, A. D. 1270–198, Ph.D. thesis, Univ. of Colo., Boulder.

Barlow, L. K., J. W. C. White, R. G. Barry, J. C. Rogers, and P. M. Grootes (1993), The North Atlantic oscillation signature in deuterium and deuterium excess signals in the Greenland Ice Sheet Project 2 ice core, 1840–1970, *Geophys. Res. Lett.*, *20*, 2901–2904.

Bigeleisen, J., M. Perlman, and H. Prosser (1952), Conversion of hydrogenic materials to hydrogen for isotopic analysis, *Anal. Chem.*, *24*, 1356–1357.

Boyle, E. A. (1997), Cool tropical temperatures shift the global δ^{18}O-T relationship: An explanation for the ice core δ^{18}O-borehole thermometry conflict?, *Geophys. Res. Lett.*, *24*, 273–276, doi:10.1029/97GL00081.

Bradley, R., K. Briffa, J. Cole, M. Hughes, and T. Osborn (2003), The climate of the last millennium, in *Paleoclimate, Global Change and the Future*, edited by K. Alverson, R. Bradley, and T. Pedersen, pp. 105–141, Springer, Berlin.

Ciais, P., and J. Jouzel (1994), Deuterium and oxygen 18 in precipitation: Isotopic model, including mixing cloud processes, *J. Geophys. Res.*, *99*(D8), 16,793–16,803.

Cook, E., K. Briffa, D. Meko, D. Graybill, and G. Funkhouser (1995), The "segment length curse" in long tree-ring chronology development for palaeoclimatic studies, *Holocene*, *5*, 229–237.

Coplen, T. (1988), Normalization of oxygen and hydrogen isotope data, *Chem. Geol.*, *72*, 293–297.

Cuffey, K. M., G. D. Clow, R. B. Alley, M. Stuiver, E. D. Waddington, and R. W. Saltus (1995), Large arctic temperature change at the Wisconsin-Holocene glacial transition, *Science*, *270*, 455–458.

Dansgaard, W. (1964), Stable isotopes in precipitation, *Tellus*, *16*, 436–468.

Divine, D., and C. Dick (2006), Historical variability of sea ice edge position in the Nordic Seas, *J. Geophys. Res.*, *111*, C01001, doi:10.1029/2004JC002851.

EPICA Community Members (2004), Eight glacial cycles from an Antarctic ice core, *Nature*, *429*, 623–628.

Epstein, S., and T. Mayeda (1953), Variations of the ^{18}O/^{16}O ratio in natural waters, *Geochim. Cosmochim. Acta*, *4*, 213–224.

Fischer, H., M. Werner, D. Wagenbach, M. Schwager, T. Thorsteinnson, F. Wilhelms, J. Kipfstuhl, and S. Sommer (1998), Little ice age clearly recorded in northern Greenland ice cores, *Geophys. Res. Lett.*, *25*, 1749–1752, doi:10.1029/98GL01177.

Fisher, D. A., N. Reeh, and H. B. Clausen (1985), Stratigraphic noise in the time series derived from ice cores, *Ann. Glaciol.*, *7*, 76–83.

Fritzsche, D., R. Schutt, H. Meyer, H. Miller, F. Wilhelms, T. Opel, and L. Savatyugin (2005), A 275 year ice-core record from Akademii Nauk ice cap, Severnaya Zemlya, Russian Arctic, *Ann. Glaciol.*, *42*, 361–366.

Ghil, M., et al. (2002), Advanced spectral methods for climatic time series, *Rev. Geophys.*, *40*(1), 1003, doi:10.1029/2000RG000092.

Gray, S. T., L. J. Graumlich, J. L. Betancourt, and G. T. Pederson (2004), A tree-ring based reconstruction of the Atlantic Multidecadal Oscillation since 1567 A. D., *Geophys. Res. Lett.*, *31*, L12205, doi:10.1029/2004GL019932.

Grinsted, A., J. C. Moore, V. Pohjola, T. Martma, and E. Isaksson (2006), Svalbard summer melting, continentality, and sea ice extent from the Lomonosovfonna ice core, *J. Geophys. Res.*, *111*, D07110, doi:10.1029/2005JD006494.

Grosfeld, K., G. Lohmann, N. Rimbu, K. Fraedrich, and F. Lunkeit (2007), Atmospheric multidecadal variations in the North Atlantic realm: Proxy data, observations, and atmospheric circulation model studies, *Clim. Past*, *3*, 39–50.

Hanssen-Bauer, I., and E. Førland (1998), Long-term trends in precipitation and temperature in the Norwegian Arctic: Can they be explained by changes in atmospheric circulation patterns, *Clim. Res.*, *10*, 143–153.

Helsen, M. M., R. S. W. van de Wal, M. R. van den Broeke, V. Masson-Delmotte, H. A. J. Meijer, M. P. Scheele, and M. Werner (2006), Modeling the isotopic composition of Antarctic snow using backward trajectories: Simulation of snow pit records, *J. Geophys. Res.*, *111*, D15109, doi:10.1029/2005JD006524.

Hoffmann, G., J. Jouzel, and S. Johnsen (2001), Deuterium excess record from central Greenland over the last millennium: Hints of a North Atlantic signal during the Little Ice Age, *J. Geophys. Res.*, *106*, 14,265–14,274, doi:10.1029/2000JD900585.

Isaksson, E., et al. (2001), A new ice core record from Lomonosovfonna, Svalbard: Viewing the data between 1920–1997 in relation to present climate and environmental conditions, *J. Glaciol.*, *47*(157), 335–345.

Isaksson, E., et al. (2005), Two ice-cored ^{18}O records from Svalbard illustrating climate and sea-ice variability over the last 400 years, *Holocene*, *15*, 501–509, doi:10.1191/0959683605hl820rp.

Jansen, E. et al. (2007), Palaeoclimate, in *Climate Change 2007: The Physical Science Basis—Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by S. Solomon et al., pp. 433–498, Cambridge Univ. Press, Cambridge, U. K.

Johnsen, S., W. Dansgaard, and J. White (1989), The origin of arctic precipitation under present and glacial conditions, *Tellus, Ser. B*, *41*, 452–468.

Jouzel, J., and L. Merlivat (1984), Deuterium and oxygen 18 in precipitation: Modeling of the isotopic effects during snow formation, *J. Geophys. Res.*, *89*, 11,749–11,758.

Jouzel, J., C. Genthon, C. Lorius, J. R. Petit, and N. I. Barkov (1987), Vostok ice core—A continuous isotope temperature record over the last climatic cycle (160,000 years), *Nature*, *329*, 403–408.

Jouzel, J., M. Stiévenard, S. J. Johnsen, A. Landais, V. Masson-Delmotte, A. Sveinbjornsdottir, F. Vimeux, U. von Grafenstein, and J. W. C. White (2007a), The GRIP deuterium-excess record, *Quat. Sci. Rev.*, *26*, 1–2.

Jouzel, J., et al. (2007b), Orbital and millennial Antarctic climate variability over the past 800,000 years, *Science*, *317*(5839), 793–796, doi:10.1126/science.1141038.

Juckes, M., M. Allen, K. Briffa, J. Esper, G. Hegerl, A. Moberg, T. Osborn, S. Weber, and E. Zorita (2006), Millennial temperature reconstruction intercomparison and evaluation, *Clim. Past Discuss.*, *2*, 1001–1049.

Kekonen, T., J. Moore, P. Peramaki, R. Mulvaney, E. Isaksson, V. Pohjola, and R. S. W. van de Wal (2005), The 800 year long ion record from the Lomonosovfonna (Svalbard) ice core, *J. Geophys. Res.*, *110*, D07304, doi:10.1029/2004JD005223.

Knight, J. R., R. J. Allan, C. K. Folland, M. Vellinga, and M. E. Mann (2005), A signature of persistent natural thermohaline circulation cycles in observed climate, *Geophys. Res. Lett.*, *32*, L20708, doi:10.1029/2005GL024233.

Koerner, R. M. (1997), Some comments on climatic reconstructions from ice cores drilled in areas of high melt, *J. Glaciol.*, *43*, 90–97.

Kotlyakov, V. M., S. M. Arkhipov, K. A. Henderson, and O. V. Nagornov (2004), Deep drilling of glaciers in Eurasian Arctic as a source of paleoclimatic records, *Quat. Sci. Rev.*, *23*, 1371–1390.

Krinner, G., and M. Werner (2003), Impact of precipitation seasonality changes on isotopic signals in polar ice cores: A multi-model analysis, *Earth Planet. Sci. Lett.*, *216*, 525–538, doi:10.1016/S0012-821X(03)00550-8.

Marshall, J., et al. (2001), North Atlantic climate variability: phenomena, impacts and mechanisms, *Int. J. Climatol.*, *21*, 1863–1898.

Masson-Delmotte, V., J. Jouzel, A. Landais, M. Stievenard, S. J. Johnsen, J. W. C. White, M. Werner, A. Sveinbjornsdottir, and K. Fuhrer (2005a), GRIP deuterium excess reveals rapid and orbital-scale changes in Greenland moisture origin, *Science*, *309*(5731), 118–121, doi:10.1126/science.1108575.

Masson-Delmotte, V., et al. (2005b), Holocene climatic changes in Greenland: Different deuterium excess signals at Greenland Ice Core Project (GRIP) and NorthGRIP, *J. Geophys. Res.*, *110*, D14102, doi:10.1029/2004JD005575.

Meijer, H., R. Neubert, and G. Visser (2000), Cross contamination in dual inlet isotope ratio mass spectrometers, *Int. J. Mass Spectrom.*, *198*, 45–61.

Moore, J. C., A. Grinsted, T. Kekonen, and V. Pohjola (2005), Separation of melting and environmental signals in an ice core with seasonal melt, *Geophys. Res. Lett.*, *32*, L10501, doi:10.1029/2005GL023039.

Moore, J., T. Kekonen, A. Grinsted, and E. Isaksson (2006), Sulfate source inventories from a Svalbard ice core record spanning the Industrial Revolution, *J. Geophys. Res.*, *111*, D15307, doi:10.1029/2005JD006453.

North Greenland Ice Core Project Members (2004), High-resolution record of Northern Hemisphere climate extending into the last interglacial period, *Nature*, *431*, 147–151, doi:10.1038/nature02805.

O'Dwyer, J., E. Isaksson, T. Vinje, T. Jauhiainen, J. Moore, V. Pohjola, R. Vaikmäe, and R. S. W. van de Wal (2000), Methanesulfonic acid in a Svalbard ice core as in indicator of ocean climate, *Geophys. Res. Lett.*, *27*(8), 1159–1162.

Opel, T., D. Fritzsche, H. Meyer, R. Schütt, K. Weiler, U. Ruth, F. Wilhelms, and H. Fischer (2008), 115-year ice core data from Akademii Nauk Ice Cap (Severnaya Zemlya)—A high resolution record of Eurasian Artic climate change, *J. Glaciol.*, in press.

Pohjola, V. A., T. A. Martma, H. A. J. Meijer, J. C. Moore, E. Isaksson, R. Vaikmäe, and R. S. W. van de Wal (2002a), Reconstruction of three centuries of annual accumulation rates based on the record of stable isotopes of water from Lomonosovfonna, Svalbard, *Ann. Glaciol.*, *35*, 57–62.

Pohjola, V. A., J. C. Moore, E. Isaksson, T. Jauhiainen, R. S. W. van de Wal, T. Martma, H. A. J. Meijer, and R. Vaikmäe (2002b), Effect of periodic melting on geochemical and isotopic signals in an ice core from Lomonosovfonna, Svalbard, *J. Geophys. Res.*, *107*(D4), 4036, doi:10.1029/2000JD000149.

Polyakov, I., U. Bhatt, H. Simmons, D. Walsh, J. Walsh, and X. Zhang (2005), Multidecadal variability of North Atlantic temperature and salinity during the twentieth century, *J. Clim.*, *18*, 4562–4580.

Rasmussen, L. A., and J. Kohler (2007), Mass balance of three Svalbard glaciers reconstructed back to 1948, *Polar Res.*, *26*(2), 168–174, doi:10.1111/j.1751-8369.2007.00023.x.

Serreze, M. C., F. Carse, R. G. Barry, and J. C. Rogers (1997), Icelandic low cyclone activity: Climatological features, linkages with the NAO, and relationships with recent changes in the Northern Hemisphere circulation, *J. Clim.*, *10*, 453–464.

Smith, T., and R. Reynolds (2004), Improved extended reconstruction of SST (1854–1997), *J. Clim.*, *17*, 2466–2477.

Stahrenberg, B. (1985), CO_2/H_2O equilibration device for $^{18}O/^{16}O$, *Tech. Rep. 421*, Finnigan-MAT, Bremen, Germany.

Stenni, B., V. Masson-Delmotte, S. Johnsen, J. Jouzel, A. Longinelli, E. Monnin, R. Röthlisberger, and E. Selmo (2001), An oceanic cold reversal during the last deglaciation, *Science*, *293*, 2074–2077, doi:10.1126/science.1059702.

Tarrusov, A. (1992), The Arctic from Svalbard to Severnaya Zemlya: Climatic reconstructions from ice cores, in *Climate Since A. D. 1500*, edited by R. Bradley and P. Jones, pp. 505–516, Routledge, London.

van de Wal, R. S. W., R. Mulvaney, E. Isaksson, J. C. Moore, J. F. Pinglot, V. A. Pohjola, and M. P. A. Thomassen (2002), Reconstruction of the historical temperature trend from measurements in a medium-length borehole on the Lomonosovfonna plateau, Svalbard, *Ann. Glaciol.*, *35*, 371–378.

Vimeux, F., V. Masson, J. Jouzel, M. Stievenard, and J. R. Petit (1999), Glacial-interglacial changes in ocean surface conditions in the Southern Hemisphere, *Nature*, *398*, 410–413, doi:10.1038/18860.

Vimeux, F., V. Masson, G. Delaygue, J. Jouzel, J. R. Petit, and M. Stievenard (2001), A 420,000 year deuterium excess record from East Antarctica: Information on past changes in the origin of precipitation at Vostok, *J. Geophys. Res.*, *106*, 31,863–31,874, doi:10.1029/2001JD900076.

Vinje, T. (2001), Anomalies and trends of sea-ice extent and atmospheric circulation in the nordic seas during the period 1864–1998, *J. Clim.*, *14*(3), 255–267.

Virkkunen, K., J. Moore, E. Isaksson, V. Pohjola, P. Peramaki, and T. Kekonen (2007), Warm summers and ion concentrations in snow: Comparison with present day and Medieval Warm Epoch from snowpits and an ice core from Lomonosovfonna, *Svalbard, J. Glaciol.*, *53*(183), 623–634.

Watanabe, O., H. Motoyama, M. Igarashi, K. Kamiyama, S. Matoba, K. Goto-Azuma, H. Narita, and H. Kameda (2001), Studies on climatic and environmental changes during the last few hundred years using ice cores from various sites in Nordaustlandet, Svalbard, *Mem. Natl. Inst. Polar Res.*, *54*, 227–242.

Werner, M., M. Heimann, and G. Hoffmann (2001), Isotopic composition and origin of polar precipitation in present and glacial climate simulations, *Tellus, Ser. B*, *53*, 53–71, doi:10.1034/j.1600-0889.2001.01154.x.

Whittaker, L. M., and L. H. Horn (1981), Geographical and seasonal distribution of North American cyclogenesis, 1958–1977, *Mon. Weather Rev.*, *109*, 2312–2322.

Whittaker, L. M., and L. H. Horn (1984), Northern Hemisphere extratropical cyclone activity for four mid-season months, *Int. J. Climatol.*, *4*(3), 297–310.

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