Svalbard summer melting, continentality, and sea ice extent from the Lomonosovfonna ice core

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Received 13 July 2005; revised 25 October 2005; accepted 18 January 2006; published 14 April 2006.

[1] We develop a continentality proxy (1600–1930) based on amplitudes of the annual signal in oxygen isotopes in an ice core. We show via modeling that by using 5 and 15 year average amplitudes the effects of diffusion and varying layer thickness can be minimized, such that amplitudes then reflect real seasonal changes in $\delta^{18}O$ under the influence of melt. A model of chemical fractionation in ice based on differing elution rates for pairs of ions is developed as a proxy for summer melt (1130–1990). The best pairs are sodium with magnesium and potassium with chloride. The continentality and melt proxies are validated against twentieth-century instrumental records and longer historical climate proxies. In addition to summer temperature, the melt proxy also appears to reflect sea ice extent, likely as a result of sodium chloride fractionation in the oceanic sea ice margin source area that is dependent on winter temperatures. We show that the climate history they depict is consistent with what we see from isotopic paleothermometry. Continentality was greatest during the Little Ice Age but decreased around 1870, 20–30 years before the rise in temperatures indicated by the $\delta^{18}O$ profile. The degree of summer melt was significantly larger during the period 1130–1300 than in the 1990s.

Citation: 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.

1. Introduction

[2] Ice cores hold many potential proxies of climate. The difficulty is usually to assign a proxy to a particular climatic variable [*Jones and Mann*, 2004]. One of the most well known is $\delta^{18}O$, which is the deviation of the isotopic ratio between H_2O^{18} to H_2O^{16} from the ratio in standard seawater [*Dansgaard*, 1964]. In precipitation $\delta^{18}O$ is generally accepted as being a proxy for condensation temperature in the atmosphere and is related to the surface temperature in modern snow [*van Lipzig et al.*, 2002].

[3] However, the relationship between $\delta^{18}O$ and surface temperature changes under different climate regimes as it is affected by changes in moisture source temperature, seasonality of precipitation and moisture transport path. In addition to mean temperature, the seasonal difference between summer and winter is of broad environmental significance. Continental interiors have larger seasonal temperature variations than maritime areas and the annual temperature range is often used as a measure of continentality. Ice cores from the interiors of large ice sheets such as Greenland or Antarctica are probably not very sensitive to the location of the sea ice margin, which determines the distance to open water and hence influences continentality, as it is far away from the drilling site. However, an ice core from an island near the seasonal sea ice margin may reflect the much larger relative variations in distance to open water via a continentality index. In ice cores where the seasonal cycle in $\delta^{18}O$ is well preserved the amplitude is an obvious candidate as a proxy for continentality. Unfortunately the oxygen isotope record is often affected by diffusion on a scale comparable to the annual layer thickness. It is relatively common to back diffuse the isotopic profile to amplify the annual cycle for layer counting purposes [*Bolzan and Pohjola*, 2000]. However, the reconstructed amplitude from back diffusion is very sensitive to small errors [*Johnsen et al.*, 2000] and, in sites with seasonal melt, back diffusion is problematic as it is expected that diffusion acts very unevenly.

[4] Stratigraphic melt index (SMI) is defined as the weight fraction of clear bubble-free ice. SMI has been used as a proxy for summer temperature on Canadian ice caps [e.g., *Koerner*, 1997]. However, in Greenland, *Pfeffer and Humphrey* [1998] suggest that melt layers not only reflect warm summer conditions but also the temperature gradient experienced by the snowpack during the preceding months. Therefore, under some conditions, SMI is also affected by the seasonal temperature range, or continentality, and accumulation rate.

[5] Almost all ice cores in the Arctic outside of Greenland are influenced by seasonal melt. Meltwater can refreeze on the surface, percolate into deeper layers or run off. If meltwater refreezes in the surface layers, then it has little impact on average ion concentrations and $\delta^{18}O$ values.

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0148-0227/06/2005JD006494$09.00

Percolation and runoff affects ions much more strongly than isotopes because most ions have a very strong affinity for the water phase and even small amounts of meltwater will have high concentrations of ions [*Davies et al.*, 1982]. Percolation however only redistributes the ions in the snowpack whereas they are lost by runoff. *Iizuka et al.* [2002] use the preferential removal of ions from particular layers as an indicator for seasonal melt on Austfonna, Svalbard, and find empirically that the magnesium sodium ratio was the best melt proxy.

[6] In 1997 we drilled a 121-m-long ice core (spanning about 800 years) on Lomonosovfonna [*Isaksson et al.*, 2001], the highest ice field in Svalbard (78°51′53″N, 17°25′30″E, 1255 m above sea level (asl)). 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. Dating of the core was based on a layer-thinning model tied with the known dates of prominent reference horizons (see *Kekonen et al.* [2005] for details). There is evidence that $\delta^{18}O$ is a good proxy for surface air temperature at the drill site: the rise in $\delta^{18}O$ from 1900 to 1920 that mirrors the rise in temperature seen in the instrumental record from Svalbard [*Isaksson et al.*, 2003], consistency with the borehole temperature profile [*van de Wal et al.*, 2002], and well-preserved annual cycles [*Pohjola et al.*, 2002b]. The current annual temperature range is from 0°C to about −40°C, though temperatures within the firn pack tend to be much warmer, approaching −3°C at 15 m depth [*van de Wal et al.*, 2002].

[7] The soluble ion chemical record in the ice core spans the last 800 years [*Kekonen et al.*, 2005] and shows clearly defined changes corresponding with different climatic periods and anthropogenic pollution. For the Lomonosovfonna core the impact of percolation is much less significant than the impact of changes in precipitation chemistry [*Moore et al.*, 2005]. All ions except ammonium have anomalously low concentrations in the oldest ice of the core. This is interpreted as loss of ions in the original snowpack caused by warm conditions prior to about 1300 [*Kekonen et al.*, 2005]. SMI for the core shows no relationship with oxygen isotopes [*Pohjola et al.*, 2002a], and thus we cannot interpret SMI as a proxy for summer temperature at the site.

[8] In this paper we develop the theoretical basis for both continentality and melt proxies and apply them to the chemical [*Kekonen et al.*, 2005] and the oxygen isotope records [*Isaksson et al.*, 2003] from the Lomonosovfonna ice core. We test the ice core proxies against instrumental temperature and historical sea ice extent over the last 150 years and then use the ice core proxies to derive climatic variables back to 1130. While summer temperature proxies are common, such as tree ring width chronologies, we will show that we have proxies of both summer melting and annual temperature range, both of which seem to be influenced by regional sea ice extent.

2. Data

[9] The oxygen isotopic record ($\delta^{18}O$) has been measured at 5 cm resolution [*Isaksson et al.*, 2003]. The concentration of ions (Cl^-, NO_3^-, SO_4^{2-}, $CH_3SO_3^-$, NH_4^+, K^+, Ca^{2+}, Mg^{2+}, Na^+) has been measured in 5 cm samples taken at 10 cm intervals along the whole core. Sample recovery and analysis are discussed by *Kekonen et al.* [2005, and references therein]. SMI has been measured at 5 mm resolution in the top 80 m (1707−1997 A.D.) of the core by visually classifying into firn, bubbly ice, diffuse ice and clear ice and assigning a melt percentage to each category [*Pohjola et al.*, 2002a].

[10] We will compare our ice core proxies to local historical temperatures and also to Barents Sea ice extent since this is a major driving force of regional climate [*Isaksson et al.*, 2003, 2005a, 2005b]. Svalbard Airport reconstructed temperature, from 80km south west of the drill site, is the closest instrumental historical record. This is a composite record from sites near Longyearbyen and has monthly mean temperatures since 1911 [*Nordli et al.*, 1996]. We extract winter (T_{DJF}), summer (T_{JJA}) and yearly mean temperatures (T_m) as well as yearly temperature range (ΔT). Records of Barents sea ice extent have been compiled from ship logbooks [*Vinje*, 2001]. April sea ice extent for the eastern and western Barents Sea (B_e and B_w, respectively) covers the period from 1864 to 1999, with few gaps most notably in the WWII years. We have left the missing years out of correlations with other records.

3. Methods

3.1. Continentality From $\delta^{18}O$

3.1.1. Factors Contributing to $\delta^{18}O$ Amplitudes

[11] The amplitude of the annual signal in $\delta^{18}O$ (A) is an obvious candidate for a continentality proxy, since $\delta^{18}O$ on Lomonosovfonna is a proxy for surface air temperature [*Isaksson et al.*, 2001; *van de Wal et al.*, 2002] and the annual cycle is well preserved [*Pohjola et al.*, 2002b]. However, there are complicating factors in using the measured amplitudes. Meltwater percolation redistributes isotopes, effectively reducing the annual amplitudes. Further, the amplitudes decay with depth because of isotopic diffusion [*Johnsen et al.*, 2000].

[12] Diffusion acts so as to smooth out steep gradients in the signal, so thin layers with a given initial amplitude of $\delta^{18}O$ are smoothed faster than thick layers with the same initial amplitude. Therefore one could argue that A is a proxy for accumulation rate rather than for continentality. In the firn layer (top 30 m on Lomonosovfonna) diffusion is very fast because of vapor transport, and therefore amplitudes decay rapidly in this layer. Solid ice diffusion is a much slower process than firn diffusion. The modeled amplitudes would change only about 5% between the bottom of the firn layer (defined as the depth were pores are closed off from the atmosphere) and AD1600 (using diffusion rates from *Johnsen et al.* [2000]), and so solid ice diffusion can be ignored. Thin ice layers formed during melt/refreeze events will act as vapor barriers and limit firn diffusion. To ensure we are below the effect of vapor transport, we restrict ourselves to the record below 30 m. That means we can reasonably assume that there is very little trend in the characteristic diffusion length scale in this part of the core, though there will be local variability (e.g., due to ice layers).

[13] Previous studies on Lomonosovfonna have found that percolation lengths are generally shorter than the annual layer thickness [*Pohjola et al.*, 2002a] except during the warmest years (such as post-1990) where percolation length

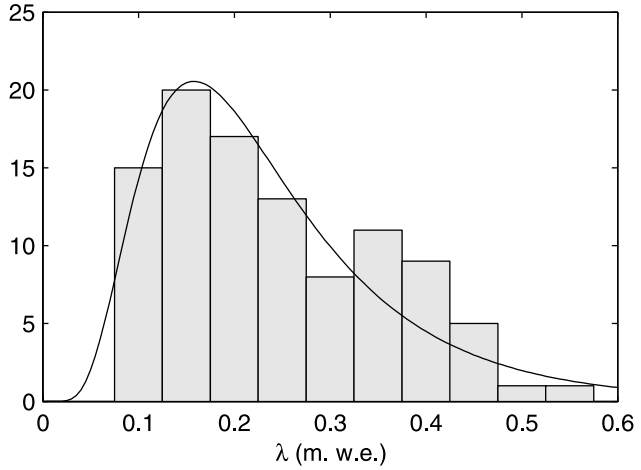

Figure 1. Histogram of layer thickness from 25 to 37 m (i.e., near pore close off) and the lognormal fit used in the modeling.

appear to be 2–8 annual layers [*Kekonen et al.*, 2005]. From snow pit studies carried out in May 2002 on Lomonosovfonna we find that the annual isotopic range prior to melt is approximately 10‰. We consider now the scenario where a fraction of the summer snowmelts, percolates and then refreezes in the winter snow. Prior to melt we assume that $\delta^{18}O_{summer} = -10‰$ and $\delta^{18}O_{winter} = -20‰$. Isotopic fractionation is observed to occur during freezing of ice, where an isotopic gradient will form in the ice layer because the first ice that forms will be relatively enriched in $H_2^{18}O$ because of Rayleigh fractionation [e.g., *Hubbard and Sharp*, 1993]. The ice will be ~3‰ heavier than the water it freezes from. The same effect is not seen during melting and the meltwater has the same isotopic content as the summer snow ($\delta^{18}O_{water} = -10‰$) because the ice in individual crystals is effectively unmixed. The refrozen meltwater will therefore have average $\delta^{18}O$ of −10‰ with the heaviest fraction having −7‰. However, it will have been mixed with a fraction of winter snow. In the case where the refrozen layer composition originates from one third meltwater and two thirds winter snow (equivalent to ~50% SMI) the resulting layer will have mean $\delta^{18}O_{refrozen} = -16.7‰$ with the heaviest being −15.7‰. The net effect is that the annual isotopic range will have been reduced from 10‰ to ~6‰ because of percolation, which compares well with the isotopic values in the upper firn pack on Lomonosovfonna [*Pohjola et al.*, 2002a]. However, thin ice layers will not cause any significant smoothing and could even inhibit firn diffusion. For this reason we may expect that the amplitudes respond nonlinearly to SMI. Surface melt will generally dampen the signal whereas limited melt inhibits smoothing by firn diffusion. Observational data (Figure 4) will be discussed later.

3.1.2. Accumulation Rate Effects on Amplitudes

[14] Rather than attempting the difficult back diffusion problem, which is sensitive to small errors in diffusion length, we argue that a statistical mean over several years will minimize the integrated effects of diffusion. That is, we will assume that adequate time smoothing will give a record that is close to proportional to the annual amplitude in freshly precipitated snow (A_0). To test this assumption, we will apply a mathematical diffusion model to an artificial record of amplitudes with realistically distributed layer thickness. From this we will give an estimate of how much of the total variance in A is due to diffusion processes and so how much smoothing is needed. We define σ as the characteristic diffusion length scale [*Johnsen et al.*, 2000]. The observed layer thickness (λ) has a wide distribution (Figure 1), varying by a factor of 5 or more, while variations in σ due to ice layers are, we argue, much smaller; we will return to this point later. The ratio λ/σ, which measures the annual layer scale to the diffusion length scale, will be dominated by changes in λ and we will assume a constant σ in the development that follows.

[15] The net effect of diffusion is equivalent to convoluting the deposited signal with a Gaussian filter having a standard deviation σ (which is also the diffusion length), [*Johnsen et al.*, 2000]. Using the Fourier convolution theorem, the power spectral density of the isotopic profile is then

$$P(F) = P_0(F)e^{-(2\pi F\sigma)^2}, \tag{1}$$

where F is the frequency and P_0 is the spectral density of the initial deposited $\delta^{18}O$ profile, which is, however, compressed by layer thinning with depth due to ice flow. We will estimate P_0 from the near surface layers and then estimate σ by fitting equation (1) to the measured P from a section of the profile just below pore close off. If we assume that the isotopes are deposited with a red noise (AR1) background spectrum with lag-1 autocorrelation α, we can write

$$P_{surf}(F) = 2Vdz\frac{1-\alpha^2}{\left|1-\alpha e^{-2i\pi Fdz}\right|^2}, \tag{2}$$

where dz is the sampling resolution and V is the variance of the deposited signal [*Torrence and Compo*, 1998]. We will use water equivalent (w.e.) depth units to minimize layer thinning with depth, and therefore be less sensitive to precise determination of pore close off depth. We estimate the deposited signal to have $\alpha = 0.85$ from the top 4 m w.e. of the core, sampled at a resolution of 0.03 m w.e. At depth, the background spectrum will have been both whitened because of compression and reddened because of diffusion. We assume that P_0 is proportional to P_{surf} with the frequency axis stretched to account for thinner layers at depth. We can then write $P_0(F) = cP_{surf}(rF)$, where r accounts for spectrum whitening and c represents the difference in isotopic variance between the present and at depth. We determine the diffusion length by fitting equation (1) (with an additional white measurement noise term) to the observed power spectrum using the maximum entropy method [*Chen*, 1988]. The measurement noise is estimated to have a standard deviation of 0.06‰ from the high-frequency variability of the oldest and most diffused ice. This is consistent with an estimated measurement precision of ±0.1‰ based on analysis of replica samples. In Figure 2 the diffusion length has been determined through minimizing the mean absolute deviation of the theoretical spectrum to the observed in a moving window and has then been decompressed to pore close off level. The variation in diffusion length observed in

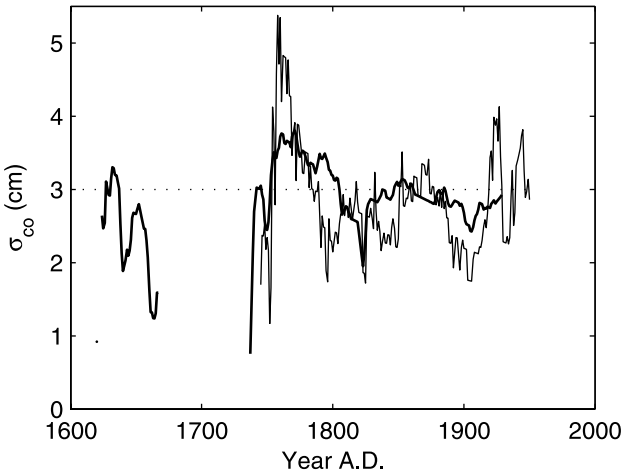

Figure 2. Best fitting diffusion length decompressed to pore close-off level, obtained by minimizing the mean absolute deviation between the theoretical power spectrum (equation (1)) and the low-order maximum entropy power spectrum estimated in 15 year (thin line) and 60 year (thick line) windows. The curve based on 15 year windows is cut off at 1740 because there are too few samples in the window to accurately estimate the power spectrum. Similarly, the gap near 1700 is due to discontinuous sampling. The variability in σ_{co} increases back in time as the number of samples in the window decreases and as compression makes it problematic to resolve the diffusion length with the given sampling resolution. This suggests that the variability reflects fitting uncertainty rather than a climate signal.

Figure 2 increases back in time and does not correlate with SMI nor is there any hint of a nonlinear relationship in a scatterplot. Therefore the variation is more likely caused by uncertainties in the estimated spectrum due to too few samples in the short window or because compression makes it harder to resolve the diffusion length with the given sampling resolution, rather than actual variations in diffusion length. The fact that there is practically no trend in the diffusion length provides justification for the assumption that solid ice diffusion can be ignored.

[16] As a consequence of equation (1), the amplitude of the annual signal can be written as

$$A = A_0 e^{-2\left(\pi\frac{\sigma}{\lambda}\right)^2}, \tag{3}$$

where A_0 is the initial amplitude and λ is the compressed layer thickness. Measured λ [*Pohjola et al.*, 2002b] is close to lognormally distributed (Figure 1). The $\log_{10}(\lambda)$ has mean -0.67 and standard deviation 0.22 at pore close off. Using these statistics, we generate a long surrogate time series of λ and calculate how a diffusion length $\sigma = 0.030$ m w.e. affects the amplitudes. The result can be seen in Figure 3. At 30 m depth, A has on average decayed to 61% of the deposited A_0, but its distribution is very wide. Table 1 shows that 5% of the layers will have decayed to less than 13% of the initial value, while another 5% of layers only decayed by 7%. However, taking 15 year moving averages of A tightens the distribution significantly (Figure 3). The proportion of variance (W) in A due to variation in λ can be estimated as

$$W = \frac{\overline{A}^2 \mathrm{var}\left(e^{-2\left(\pi\frac{\sigma}{\lambda}\right)^2}\right)}{\mathrm{var}(A)}, \tag{4}$$

where $\overline{A}$ is the mean value of A around pore close off. Using the variances from Table 1, we find that with 5 and 15 year moving averages only 11% and 7% of the total variance in A is due to variations in λ, so we can say that A is nearly proportional to A_0. That is, adequate time smoothing will give a statistically representative sampling of both thin and thick layers. In a similar fashion we determine that a 25% change in σ causes a $\sim$10% change in amplitudes. Our hypothesis is therefore that smoothed A on Lomonosovfonna reflects real seasonal changes in δ^{18}O influenced by melt rather than changes in accumulation rate or diffusion length. We will test this hypothesis in the results section.

3.1.3. Amplitude Reconstruction

[17] To derive the annual amplitudes record, we first upsampled the isotope record to 10 samples per year, and then high-pass filtered it with a cutoff wavelength of 3 years. The amplitude was then measured as the range in a window 1 year wide from the modeled timescale [*Kekonen et al.*, 2005]. This processing can be carried out beyond the limits of layer counting because measuring the amplitude is substantially easier than counting the layers. Even if this procedure misses a peak and picks the one from a neighboring year, the amplitude will still be a reasonable estimate. We tested our results by varying the window length and the high-pass-filtering method, and found them to be robust.

[18] The amplitudes decay as the yearly layer thickness is compressed by ice flow, eventually to a thickness comparable with the 5 cm sampling resolution of δ^{18}O. The effect of the sampling is equivalent to convoluting the "true"

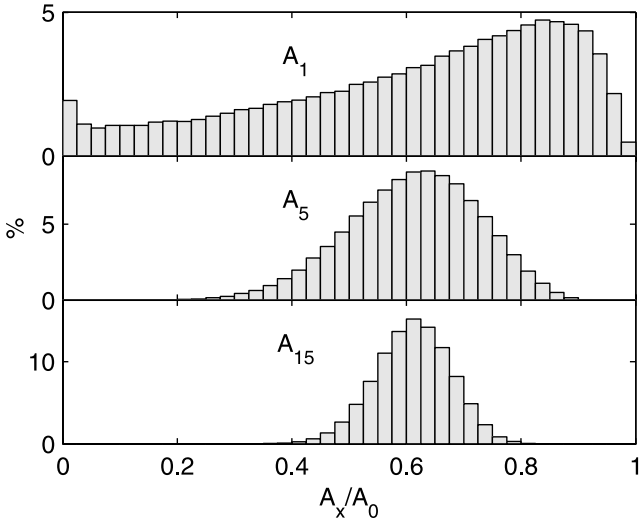

Figure 3. (top) Histogram of modeled annual amplitudes at pore close off relative to deposited amplitudes assuming a diffusion length of 3.0 cm w.e. (middle and bottom) Using 5 and 15 year moving averages (middle and bottom) tightens and normalizes the distribution significantly.

Table 1. Percentiles and Variance of Modeled Annual Isotopic Amplitudes (Equation (3)) After Firn Diffusion Relative to Their Deposited Amplitudes Assuming That the Layer Thickness is Lognormally Distributed[a]

	Percentile			
	5%	50%	95%	Variance
A/A_0	0.12	0.67	0.93	0.066
A_5/A_0	0.42	0.62	0.78	0.013
A_{15}/A_0	0.51	0.62	0.72	0.005

[a] A_5 and A_{15} denote 5 and 15 year moving averages, respectively.

signal with a rectangle function. Hence, using the Fourier convolution theorem, we correct the amplitudes by applying

$$A_{corrected} = \frac{A_{measured}}{\mathrm{sinc}\left(\pi \frac{\Delta z}{\lambda}\right)} \quad (5)$$

where Δz is the width of a sample and λ is the local wavelength of the annual signal. At 1700 the correction factor is 1.3, while at 1600 the factor we apply is 1.6. Somewhat arbitrarily, we decide that beyond 1600 the errors in the correction factor are too big to extract a meaningful signal.

3.2. Chemical Washout Index

[19] Many studies have found that particular ions are preferentially washed out of the ice during seasonal melt [e.g., *Davies et al.*, 1982; *Iizuka et al.*, 2002]. However, details of specific ion elution rates depend on the incorporation of ions in the ice during grain growth and so are certainly affected by processes that occur prior to the onset of melting [*Cragin et al.*, 1996]. While microphysical processes governing the removal of ions are poorly understood, it is reasonable to assume that the ions are removed fractionally as the melt progresses, and so we can write

$$\frac{\partial C}{\partial W} = -a \cdot C \text{ and } C(0) = C_p \Leftrightarrow C(W) = C_p \cdot e^{-aW}, \quad (6)$$

where W is a dimensionless washout index, C is the concentration of a chemical species, a is a species-specific dimensionless washout rate and C_p is the species initial concentration in the fresh snow. Here we have implicitly assumed that the reservoir volume for the ions is constant; that is, the increase in concentration due to a reduced volume caused by runoff is negligible, which is different from the situation in a seasonal snowpack.

[20] Some chemical ions have the same principle source. For example, both Na^+ and Mg^{2+} are mainly of sea salt origin in Svalbard snow [*Virkkunen*, 2004; *Iizuka et al.*, 2002], and so their concentrations are often taken to be linearly dependent. If we define X and Y to be the concentrations of two such chemical species, then $X_p = k_{xy}Y_p$, where X_p and Y_p are the concentrations of X and Y prior to melt, and k_{XY} is their ratio in fresh snow. From equation (6) it is clear that the logarithmic ratio of the concentrations after washout is linear in W.

$$\log\left(\frac{X}{Y}\right) = \log\left(\frac{X_p \cdot e^{-a_X W}}{Y_p \cdot e^{-a_Y W}}\right) = \log(k_{XY}) + (a_Y - a_X)W. \quad (7)$$

[21] Hence we can derive a washout index provided that the washout rates a_X and a_Y differ significantly. Note that equation (7) holds even for a finite reservoir that is depleted during the melt season, as happens with significant runoff. *Iizuka et al.* [2002] show that there are high correlations (>0.95) between the concentrations of Na^+, Cl^-, K^+ and Mg^{2+} in Svalbard dry snow from Austfonna. *Virkkunen* [2004] find similar results from Lomonosovfonna, indicating that these ions have the same (marine) source for Svalbard, justifying the use of equation (7). However, all these correlations break down in wet snow, except that between Na^+ and Cl^-. Hence the washout rates for Na^+, K^+ and Mg^{2+} must differ widely, whereas they are the similar for Na^+ and Cl^-. In particular, *Iizuka et al.* [2002] find that Mg^{2+} is washed out more efficiently than Na^+ ($a_{Mg} > a_{Na}$). Because of these differences in washout rates, we can construct two independent chemical melt indices W_{NaMg} and W_{ClK} based on $\log(Na^+/Mg^{2+})$ and $\log(Cl^-/K^+)$. We consider W_{NaMg} to be superior to W_{ClK} since relative measurement errors in K^+ are much larger than in Mg^{2+} [*Kekonen et al.*, 2002] and further K^+ has a tendency to have large peaks that are seemingly unrelated to other ions in high-resolution pit data [*Iizuka et al.*, 2002]. Using seawater ratios in $\mu g L^{-1}$, we get $k_{NaMg} = 8.3$ and $k_{ClK} = 49$. To avoid difficulties with percolation between layers [*Pohjola et al.*, 2002a; *Moore et al.*, 2005], we use 15 year mean concentrations. To compare W_{NaMg} and W_{ClK}, we scale the dimensionless indices by their mean values over the twentieth century.

4. Results

[22] We test the new indices by comparison with the instrumental record and proxy data. *Isaksson et al.* [2005b] compare the $\delta^{18}O$ profile with meteorological records from around the Barents Sea with Vardø on the Norwegian coast having the best correlation. However, the Vardø temperature record is longer than that from Longyearbyen, and therefore it is expected that it will correlate better simply because it covers the large increase in temperatures at the end of the Little Ice Age (LIA). Over the period of common overlap (1911–1997), $\delta^{18}O$ correlate better with Longyearbyen than Vardø temperatures. Therefore our interpretation is that Longyearbyen temperatures are more representative of the conditions on Lomonosovfonna than those from Vardø.

[23] Correlation coefficients between 5 year moving averages of ice core data and the historical records are shown in Table 2. As expected isotopes ($\delta^{18}O$) correlate significantly with Longyearbyen temperatures (T_m, T_{DJF}, T_{JJA}). To avoid the problems with firn diffusion completely, we should discard the continentality record above pore close off (e.g., younger than 1930). However, to test A against instrumental records we need a bigger overlap and so, as a compromise, we choose to truncate A at 1970 (15.6 m depth) by when most of the firn diffusion already has taken place. The more continental the climate conditions, the greater the temperature range with winter temperatures changing the most. So, as expected, the continentality proxy (A) correlates with temperature range (ΔT), while $\delta^{18}O$ anticorrelates (Table 2). On longer timescales we find that A is also highly anticorrelated with $\delta^{18}O$ (Table 3). This is consistent with the interpretation of $\delta^{18}O$ as a temperature proxy and A as a continentality proxy. We find no clear relationship between A and SMI (Figure 4) and we

Table 2. Significant Correlation Coefficients ($p < 0.05$) Between Ice Core Data and Historical Records for Their Overlapping Periods[a]

	$\delta^{18}O$	A	SMI	W_{NaMg}	W_{ClK}
T_m	0.47	−0.38			
T_{JJA}	0.52				
T_{DJF}	0.53	−0.48			
ΔT	−0.34	0.57			
B_w	−0.65		−0.22	−0.28	−0.60
B_e	−0.34		−0.34	−0.50	−0.32

[a]All records have been smoothed using 5 year moving averages, so as to make the results insensitive to dating errors. Ice core parameters: A, continentality; SMI, stratigraphic melt index; W_{xy}, chemical washout indices. Historical parameters: T, Longyearbyen temperature (mean, summer, winter, annual range); B, April sea ice extent in the western and eastern Barents Sea.

therefore conclude that melt is not a major contributor to the variability in *A*.

[24] Linear regression gives $T_m = 1.12\ \delta^{18}O + 11.5$ ($R^2 = 0.22$) and $\Delta T = 3.3\ A + 19.5$ ($R^2 = 0.32$), and both of these regressions are significant at the 95% level. However, *van de Wal et al.* [2002] analyzed the borehole temperature profile finding that temperatures in the nineteenth century were 2.4°C colder than the twentieth century, this compares with a 0.8 ‰ difference in $\delta^{18}O$ [*Isaksson et al.*, 2003] between the nineteenth and twentieth centuries. This indicates that the above regression underestimates the sensitivity of $\delta^{18}O$ on temperature, which is to be expected considering that the historical records used in the regression only cover the relatively warm twentieth-century period [*von Storch et al.*, 2004].

[25] The SMI does not correlate significantly with temperatures. We find that the full records of SMI and $\delta^{18}O$ are significantly anticorrelated (Table 3). Thus, if we believe $\delta^{18}O$ to be a temperature proxy, it does not seem plausible to consider SMI a summer temperature proxy. The alternative explanation of SMI as a continentality index can also be rejected since there is no significant correlation with *A* (Table 3 and Figure 4). Perhaps the SMI, for this ice core, is a composite signal that requires a more sophisticated treatment of percolation effects.

[26] Perhaps surprisingly, there is no significant correlation between washout indices and any temperature series (Table 2). However, there are significant correlations with $\delta^{18}O$. These can be explained by the local differences in conditions. The summit of Lomonosovfonna is inland and 1250 m asl, whereas Longyearbyen is within a fjord at sea level. In summer, warm days often produce foggy conditions near sea level, and sunny conditions at higher elevations, therefore we expect that the washout indices are recording local summer melt.

Table 3. Significant Correlation Coefficients ($p < 0.05$) for Lomonosovfonna Ice Core Data[a]

	$\delta^{18}O$	A	SMI	W_{NaMg}	W_{ClK}
$\delta^{18}O$		−0.66	−0.48	0.60	0.51
A	−0.59			−0.46	−0.46
SMI	−0.48			−0.16	−0.12
W_{NaMg}	0.60	−0.43	−0.16		0.73
W_{ClK}	0.51	−0.36	−0.12	0.73	

[a]Parameters are as in Table 2. All records have been smoothed using 15 year moving averages.

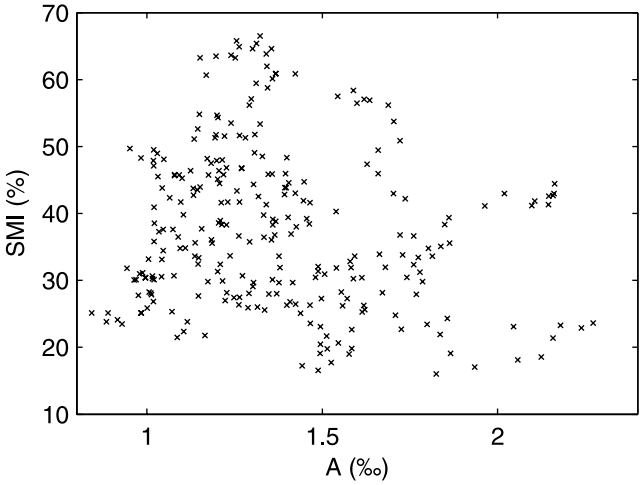

Figure 4. Scatterplot of annual amplitudes (A) against stratigraphic melt index (SMI) using 15 year moving averages (1700–1970). There is no simple relationship between melt and amplitudes. Scatter increases when using 5 year averages.

[27] It is noticeable that both washout indices have negative values throughout extensive periods (Figure 5). If, as seems reasonable, runoff is the only postdepositional process that can change 15 year averages of these ion ratios and given that runoff would act to make the indices more positive, negative values can only be explained by changing the ionic ratios of the precipitation. Either additional sources account for the excess Mg^{2+} and K^+ or fractionation processes deplete Na^+ and Cl^- relative to Mg^{2+} and K^+. Since there are simultaneous excesses of both Mg^{2+} and K^+ an additional source contributing to both ions, such as terrestrial dust (though it would need to be relatively depleted in Na^+), would produce changes in both washout rates. However, Table 2 shows that the washout rates are significantly inversely correlated with sea ice extent in the Barents Sea, which points toward fractionation processes as the most likely explanation. *Rankin et al.* [2002] proposed fractionation by frost flowers on sea ice to explain low ratios of SO_4^{2-} in Antarctic ice, but such depletion cannot be seen in the Arctic data because of the large magnitude of non-sea-salt SO_4^{2-} sources. However, the same mechanism could also remove NaCl from source aerosol. When winter temperatures drop below the NaCl eutectic (−22°C), frost flowers will be relatively enriched in ions that remain liquid at low temperatures, including Mg^{2+} and K^+. Direct evidence of depleted Na^+/Mg^{2+} and Na^+/K^+ ratios on sea ice have been found by *Domine et al.* [2004], though they suggest that an additional terrestrial source of K^+ and Mg^{2+} is involved. Two independent effects influence the ionic ratios: increased runoff raises the ratios, while increased sea ice lowers them. Thus low washout indices indicate a cold climate with relatively more sea ice. Consistently, we find that W_{NaMg} and W_{ClK} correlate with $\delta^{18}O$ and anticorrelate with *A* (Table 3).

5. Discussion

[28] The Lomonosovfonna $\delta^{18}O$ record indicates that the LIA in Svalbard ended around 1900 (Figure 5). The start of

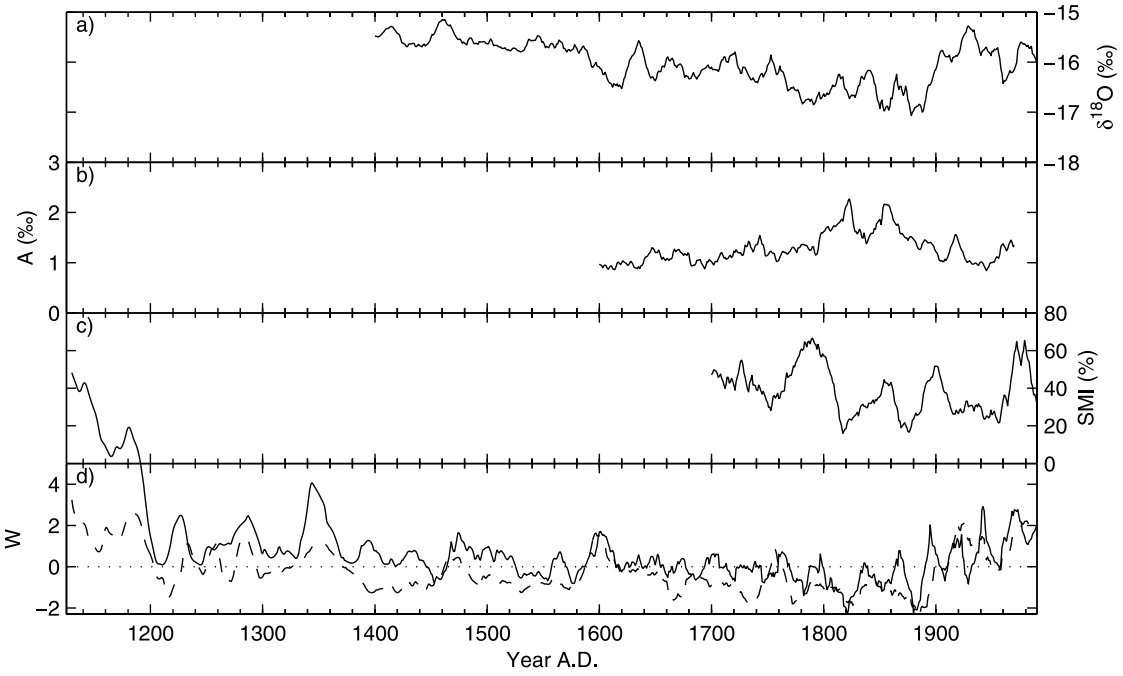

Figure 5. Fifteen-year moving averages of Lomonosovfonna ice core data. (a) Oxygen isotopes, (b) continentality proxy (A), (c) stratigraphic melt indices (SMI), and (d) washout indices (solid line is W_{NaMg}, and dashed line is W_{ClK}).

the LIA is less easy to define as there seems to be a gradual descent into colder climatic conditions. However, by 1600 a clear signature of the LIA is seen in the isotopes and chemistry [*Isaksson et al.*, 2003; *Kekonen et al.*, 2005], probably indicating that the sea ice margin had advanced close enough to Svalbard to affect local climate. Continentality and washout indices show more gradual changes. The coldest and most continental period seen in the core is the nineteenth century. For comparison, most western North American proxies also show the nineteenth century as the coldest period, whereas in European proxies the coldest period occurs between 1600 and 1700 [*Jones and Mann*, 2004]. Under the reasonable assumption that sea salt ions have the same source throughout the record, negative washout indices during the period from 1700 to 1900 are indicative of greater sea ice extent in the Barents Sea.

[29] The continentality proxy (A) peaks at 1860 and declines rapidly thereafter. This suggests that a change in climate was already setting in by 1870 and temperatures and sea ice extent lagged the continentality change by 30 years. However, continentality is most affected by winter temperatures, while sea ice extent is at a maximum in spring. Thus we suggest that winter temperatures rose several decades before spring and summer temperatures.

[30] In the oldest part of the core (1130–1200), the washout indices are more than 4 times as high as those seen during the last century, indicating a high degree of runoff. Since 1997 we have performed regular snow pit studies [*Virkkunen*, 2004], and the very warm 2001 summer resulted in similar loss of ions and washout ratios as the earliest part of the core. This suggests that the Medieval Warm Period [*Jones and Mann*, 2004] in Svalbard summer conditions were as warm (or warmer) as present-day, consistent with the Northern Hemisphere temperature reconstruction of *Moberg et al.* [2005].

6. Conclusion

[31] We hope that this paper inspires interest in using Arctic ice cores from smaller ice caps for reconstruction of Late Holocene environmental history on a regional scale. We have developed proxies for the washout intensity (W_{NaMg} and W_{ClK}) based on the ion ratios of sea salt ions (Na^+, Cl^-, Mg^{2+} and K^+) and continentality (A) based on annual amplitudes of $\delta^{18}O$. The washout indices can provide a valuable constraint on whether ion data truly represents precipitation chemistry or if runoff has affected the record. For Lomonosovfonna we find that sea ice extent also affects the washout indices, most likely because NaCl is fractionated relative to other salts in the source area. Luckily the sign of this fractionation is such that W_{NaMg} and W_{ClK} still can be thought of as proxies for summer temperatures. Washout indices confirm results from earlier studies [*Kekonen et al.*, 2005] showing that the ion chemistry in the earliest part of the core (before 1200) was dominated by washout of ions. We find that washout intensities are more than 4 times higher than the average over the last century.

[32] Annual amplitudes in $\delta^{18}O$ are affected by melt and diffusion in a nontrivial and likely nonlinear manner. In the case of Lomonosovfonna, the $\delta^{18}O$ record is subject to melt and percolation and therefore presumably also to highly anisotropic diffusion in the firn. Despite this, we show that amplitudes below the level of strongly active firn diffusion are a good proxy for continentality at Lomonosovfonna, if sufficiently smoothed. However, further studies on other cores are needed before amplitudes generally can be con-

sidered to be a proven proxy for continentality. The crucial step is to smooth the amplitude record sufficiently so that each value represents a statistical mean over several years minimizing variations in annual layer thickness. An important assumption is that the trends in diffusion length and accumulation rate are small compared with the width of the layer thickness distribution. If such trends are significant, their contribution to the total variation in A may be estimated using a straightforward modification of the model presented in this paper. Under some conditions one may argue that continentality is constant (e.g., Central Greenland over the Holocene) and A might be used to infer changes in diffusion length.

[33] On Lomonosovfonna we find that the end of the LIA is very well defined in $\delta^{18}O$ and agrees with the earliest rise in washout indices (W_{NaMg}, W_{ClK}) around 1880. The Continentality proxy (A), however, starts to decline after 1860. We suggest that the response of spring and summer temperatures and the decline in sea ice in the Barents Sea lagged the rise in winter temperatures by several decades.

[34] **Acknowledgments.** The drilling of the Lomonosovfonna 1997 ice core was financed by the Norwegian Polar Institute and the Institute for Marine and Atmospheric Research, Utrecht (IMAU), and we wish to thank the field party. Additional funding was provided by the Thule Institute, Nessling Foundation, Swedish Science Council, Norwegian Research Council, NARP, and Stiftelsen Ymer-80. Finally, the anonymous referees provided very helpful comments.

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