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Discussions

Eurasian Arctic climate over the past millennium as recorded in the Akademii Nauk ice core (Severnaya Zemlya)

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Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |

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9, 2401–2422, 2013

Eurasian Arctic climate over the past millennium

T. Opel et al.

Title Page

Abstract	Introduction
Conclusions	References
Tables	Figures

|◄ | ►|
◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Abstract

The chronology of the Akademii Nauk (AN) ice core from Severnaya Zemlya (SZ) has been expanded to the last 1100 yr. Here, we present the easternmost high-resolution ice-core climate-proxy records (δ^{18}O and sodium) from the Arctic that provide new perspectives on past climate fluctuations in the Barents and Kara seas region. Multi-annual AN δ^{18}O data as near-surface air-temperature proxy reveal major temperature changes over the last millennium, including the absolute minimum around 1800 and the exceptional warming to a double-peak maximum in the early 20th century. Neither a pronounced Medieval Climate Anomaly nor a Little Ice Age are detectable in the AN δ^{18}O record. In contrast, there is evidence for several abrupt warming and cooling events such as in the 15th and 16th centuries. These abrupt changes are probably caused by shifts in the atmospheric circulation patterns and accompanied sea-ice feedbacks in the Barents and Kara seas region that highlight the role of the internal variability of the Arctic climate system.

1 Introduction

The climate of the Arctic is characterized by ongoing drastic changes. The rapid Arctic warming is accompanied by an unprecedented sea ice retreat and other positive climate feedback processes as permafrost warming and thawing as well as melting of glaciers and ice caps. Projections of future climate change indicate a continuation and amplification of the processes in the next decades (e.g. AMAP, 2011).

Knowledge of the natural climate variability is essential to understand and assess the ongoing climate changes and for reliable predictions of future climate development. In the Arctic meteorological records are sparse and relatively short, with only very few time series starting before the 20th century (e.g. Polyakov et al., 2003b). The longest continuous time series of surface air temperature (SAT) in the Eurasian Arctic (Vardø,

Northern Norway) dates back to 1840 and in the Russian sub-Arctic the SAT time series of Arkhangelsk started in 1814 (with an interruption of two years in 1832/1833).

Hence, high-resolution climate archives such as ice cores are needed to provide sub-stantial information on the temporal and spatial patterns of the natural Arctic climate variability and its causes (e.g. Overpeck et al., 1997) as well as for an assessment of the recent Arctic warming from a centennial to millennial scale perspective. Of particu-lar interest are climate changes in the Mid- to Late Holocene, characterized by relative stable boundary conditions of the climate system and by a negligible anthropogenic influence in the pre-industrial period before about 1750 (Wanner et al., 2008).

Several paleoclimate records indicate that the recent Arctic warming and its impli-cations such as a sea-ice cover reduction are unprecedented in the Late Holocene, i.e. the last 2000 yr (e.g. Kaufman et al., 2009; Kinnard et al., 2011; Spielhagen et al., 2011). However, for a more comprehensive assessment of the natural Late Holocene climate changes in the Arctic, their causes and feedbacks, more regional paleoclimate information is required. This is particularly true for the Eurasian Arctic as compared to the North American Arctic. Neglecting Greenland and Iceland, only five out of 29 records used for the Arctic temperature reconstructions of Overpeck et al. (1997) and six out of 24 of Kaufman et al. (2009) are from the Eurasian Arctic, mostly tree-ring and lake-sediment records.

Ice-core records in the Eurasian Arctic are limited to High Arctic archipelagos such as Svalbard, Franz Josef Land (FJL) and SZ (Fig. 1). Even though glaciers and ice caps in the Eurasian Arctic are characterized by summertime surface melting and hence require special consideration of melt-water percolation processes (Koerner, 1997), they contain high-resolution paleoclimate information on a regional scale. Consequently, several ice cores were drilled in the last two decades at Svalbard and FJL (Fig. 1), providing climate records of several hundred years back in time (e.g. Henderson, 2002; Isaksson et al., 2005; Divine et al., 2011).

To extend high-resolution ice-core based paleoclimate information eastwards, a joint German-Russian team drilled a new 724 m long ice core to bedrock on the Akademii

Nauk (AN) ice cap at SZ in the Central Russian Arctic (80.52° N, 94.82° E, about 750 m a.s.l., Fig. 1) from 1999 to 2001 (Fritzsche et al., 2002). AN ice cap might be the oldest one in the Eurasian Arctic (Koerner and Fisher, 2002), even though maximum ages of 10 to 40 thousand years reported for an ice core drilled in 1986/1987 on AN ice cap are distinctly overestimated (Kotlyakov et al., 2004; Fritzsche et al., 2005). For a characterization of climate conditions as well as drilling-site characteristics including snowmelt and infiltration we refer to Opel et al. (2009). Hitherto existing results indicate that despite melt-water infiltration the AN ice core might be a key archive for the reconstruction of the Late Holocene climate and environmental history of the Eurasian Arctic (Fritzsche et al., 2005; Weiler et al., 2005; Opel et al., 2009).

In this paper, we present stable-isotope and major-ion records from the upper 411 m of AN ice core, covering the last 1100 yr. These are the first high-resolution and easternmost ice-core records from the Central Eurasian Arctic exceeding the past millennium. Therefore, they add valuable new information on the temporal and spatial patterns of climate variations in the poorly studied Eurasian Arctic. We show that this region experienced major climate variability over the past millennium and discuss long-term trend and abrupt changes as well as possible causes in an Arctic context.

2 Methods and data

The ice-core processing and sampling took place in the cold laboratory of the Alfred Wegener Institute Helmholtz Centre for Polar and Marine Research (AWI) Bremerhaven (for details see Fritzsche et al., 2005; Opel et al., 2009).

The oxygen (δ^{18}O) and hydrogen (δD) stable-isotope compositions were analyzed in a sub-annual resolution of 2.5 cm (about 16 200 samples used for this paper) at the AWI Potsdam using a Finnigan MAT Delta-S mass spectrometer with an analytical precision of better than ±0.1‰ for δ^{18}O and ±0.8‰ for δD (Meyer et al., 2000).

The major-ion concentrations were determined for screening (single core segments, i.e. 0.3 to 1.0 m) as well as high-resolution (about 5 cm) samples for selected core

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

sections using a Dionex IC20 ion chromatograph at the glaciochemistry laboratory of AWI Bremerhaven (for details see Weiler et al., 2005).

3 Ice-core dating

The dating of the upper section (0–411 m) of AN ice core is based on reference horizons and annual layer counting. As reference horizons we used the 1963 ^{137}Cs peak caused by the fallout from nuclear bomb tests (Fritzsche et al., 2002; Pinglot et al., 2003) as well as volcanic signals. We clearly identified outstanding peaks in non-sea-salt sulfate and linked them to the major volcanic eruptions of Bezymianny (Kamchatka) in 1956, Katmai/Novarupta (Alaska) in 1912, Laki (Iceland) in 1783 and the unknown volcano in 1259 (Fig. 2), partly also detected in ice cores from Greenland (Zielinski et al., 1994), Svalbard (e.g. Moore et al., 2012) and FJL (Henderson, 2002). Additionally we interpreted a smaller peak as the imprint of the Eldgja (Iceland) eruption in 934, also found in Greenland ice cores (Zielinski et al., 1994).

For an independent annual-layer counting we used the seasonal signals of the high-resolution stable-water isotope data, which are clearly detectable in most cases, even though they might be smoothed and altered by melt-water infiltration. We differentiate single years using δ^{18}O and δD winter minima and according winter maxima in deuterium excess d, that does not show significant phase lags.

The annual-layer based chronology was then cross-checked with the reference horizons. Both approaches were combined and matched, resulting in our AN ice-core chronology (AN 2012). It comprises about 1100 yr for the analyzed core section (0–411 m), i.e. the time period 900-1998. Based on the cross-checking of both dating approaches and the comparison with an adjusted Nye model (see below) we estimate a dating uncertainty better than ±3 yr after 1783 and better than ±5 yr throughout the time period 934 to 1783.

The resulting age is distinctly lower than calculated with a standard Nye (1963) model (about 1250 yr, Fig. 3) as well as it has been claimed for the corresponding section of

the AN ice core taken in 1986/1987 (about 2700 yr) by Kotlyakov et al. (1990). Both age models underestimate the annual layer thickness and overestimate, therefore, the age (Fig. 3). Consequently, AN ice cap is much younger than previously assumed. It is not in dynamic steady state and most probably has been growing until recent times.

5 This has been tested by an adjusted Nye model that takes into account the growing ice cap by adding a growth term (Fritzsche et al., 2010). Our preliminary core chronology for the whole AN ice core based on this approach exhibits an age of about 3100 yr at a possible discordance at a depth of about 694 m, 30 m above bedrock. The lowermost core section contains ice that might be a remnant of an older, but post-glacial ice-cap

10 stage.

We calculated annual values by averaging all data between the defined annual marks (high-resolution stable water isotope data) as well as by resampling of screening data (major ions) using a polynomial fit function based on our core chronology. To minimize the effects of melt-water infiltration as well as dating uncertainties we only use running

15 mean values over five and 15 yr (5 yrm and 15 yrm, respectively) for climatic interpretation.

4   Results and discussion

4.1   Comparison to meteorological data (1800–1998)

We compared our AN δ^{18}O dataset to the two longest meteorological time series of the

20 Eurasian Arctic and sub-Arctic, i.e. Vardø and Arkhangelsk (Brohan et al., 2006) as well as to two SAT compilations based on meteorological station data (Fig. 4). Before the calculation of correlation coefficients, we detrended the time series for the respective common time periods by subtracting the linear trends. As those mainly represent the general warming in the time period of overlap, the resulting correlation coefficients are

25 distinctly lower (Table 1).

For station time series a striking similarity and the best correlation to our AN δ^{18}O record was found for Vardø in Northern Norway (1840–1998). Annual mean values (5 yrm) are strongly correlated with $r_{5yrm} = 0.55$ (Table 1). The correlation to the Arkhangelsk annual mean SAT record (1814–1998) is only slightly lower ($r_{5yrm} = 0.51$, Table 1). The best accordance to the Vardø SAT time series is also valid on a seasonal scale with correlation coefficients for autumn (SON, $r_{5yrm} = 0.51$), winter (DJF, $r_{5yrm} = 0.46$) and summer (JJA, $r_{5yrm} = 0.43$) seasons. This corresponds well with the generally year-round distribution of precipitation on AN ice cap (Opel et al., 2009).

Comparable strong correlations (Table 1; Fig. 4) are also found for the SAT compilation for the North Atlantic-Arctic boundary region (T_{NA}; 1802–1998, $r_{5yrm} = 0.47$) of Wood et al. (2010) as well as the Arctic-wide SAT anomalies compilation (1875–1998, $r_{5yrm} = 0.72$) of Polyakov et al. (2003b).

The correlations of AN δ^{18}O to Vardø and Arkhangelsk SAT presented here are slightly weaker those reported for 1883–1998 (Opel et al., 2009), indicating greater differences in the mid-19th century and/or different characteristics of the subtracted linear trends due to varying time periods. However, given the distances from the ice cap to the stations, the different altitude levels and climate regimes, the similarities and correlations presented here and for shorter time periods by Opel et al. (2009) demonstrate that despite melt-water infiltration AN δ^{18}O data can be used as reliable proxy for annual SAT in the Western Eurasian Arctic. Moreover, it indicates the strong Atlantic influence on the climate of this area, i.e. the Barents and Kara seas region. Nevertheless, based on the standardized data (Fig. 4) our AN δ^{18}O record shows a larger SAT range compared to Vardø and Arkhangelsk, reflecting the more continental and less maritime climate of SZ.

Our AN δ^{18}O data reveal a general warming trend over the 19th and 20th century (about 2‰ per century; Fig. 4). Superimposed on this trend is the early-twentieth century warming (ETCW), a major Arctic-wide climatic fluctuation between 1920 and 1940 (Wood and Overland, 2010). This event is commonly related to internal dynamics of the Arctic climate system, i.e. a stronger than normal meridional circulation and

corresponding heat transport into the European Arctic connected to a strengthened Icelandic Low and a westward expansion of the Siberian High (Grant et al., 2009; Wood and Overland, 2010). The resulting warming induced a reduction of sea-ice cover and albedo mostly in the Barents Sea region leading to a reinforced cyclonic circulation and, thus, a further advection of warm southerly air, amplifying the warming (Bengtsson et al., 2004; Crespin et al., 2009).

Our AN ice-core data show that these processes are not restricted to the European Arctic but also affected the Kara Sea region, even though with a peculiarity. In the AN δ^{18}O record, the ETCW exhibits a double-peak shape with two distinct maxima around 1921/1922 and 1937/1938 (Fig. 4), indicating two major warming pulses. This specific ETCW pattern and particularly the strong warming around 1920 are to our knowledge only detected in very few regional SAT time series (i.e. Svalbard, Vardø and Archangelsk; Fig. 4) and represent, thus, a peculiarity of the Barents and Kara seas region. AN ice core maximum δ^{18}O values during the ETCW were not reached again in the 20th century and, moreover, represent the highest of the entire AN ice core record.

Additionally, the first AN δ^{18}O maximum around 1921/1922 lags almost one decade behind an unprecedented rise in sodium concentration (Fig. 5). As sodium is generated predominantly from sea salt this indicates major shifts in the regional sea-ice dynamics and/or air-masses advection prior to the ETCW. Correspondingly, the August sea-ice extent of the Kara Sea shows a strong interannual variability from 1911 to 1915 (Polyakov et al., 2003a) that might be related to this sea-salt ion peak, whereas positive sea-ice extent anomalies are reported for the eastern Nordic Seas (Vinje, 2001) as well as the Barents Sea (Vinje, 1999), respectively.

4.2 Long-term scale climate interpretation (900–1998)

Our long-term AN δ^{18}O record (i.e. 15 yrm values) shows a slight decreasing trend (about −0.11‰ per century) from 900 to 1760, characterized by a marked decadal-scale variability (Fig. 5). This is consistent with the general Late Holocene cooling in the Arctic related to the decreasing summer insolation (Kaufman et al., 2009). However,

Eurasian Arctic climate over the past millennium

T. Opel et al.

a part of this AN δ^{18}O decrease might also reflect the growth of AN ice cap with lower AN δ^{18}O values in higher elevations due to the altitude effect. The corresponding height increase of the ice-cap surface is detectable from the long-term decreasing sea-salt (sodium) concentrations, too (Fig. 5). At 1760 the δ^{18}O drops rapidly to the lowest val-
5 ues of the entire AN ice-core record: a cold period of about one century with minimum temperatures around 1800 (Fig. 5), right after the eruption of Laki in 1783 that might have caused a distinct cooling. Moreover, this period coincides with the Dalton mini-mum of solar activity. After 1800, the AN δ^{18}O derived temperature starts to increase to the absolute maximum during the ETCW. The AN sodium record exhibits distinct
10 similarities to the AN δ^{18}O over most of the record. Generally, warmer (colder) peri-ods exhibit higher (lower) sodium concentrations (Fig. 5), indicating a direct regional link between temperature and sea-ice dynamics and/or air mass advection. Less sea-ice cover and more open water in warmer times might lead to a higher proportion of regional moisture with high sea-salt ion concentrations due to the sea-spray effect.
15   The spatial significance of our AN δ^{18}O record as SAT proxy for the Western Eurasian Arctic is underlined by the marked similarities to the Austfonna (Svalbard) δ^{18}O record (Isaksson et al., 2005) for the last six centuries as well as to the Vetreniy ice cap (FJL) δ^{18}O record (Henderson, 2002) for the last about three centuries (Fig. 5). Both cores were taken at relative low altitudes (750 and 500 m a.s.l., respectively) and
20 are, thus, comparable to AN ice core. Both records are about 3‰ less depleted in δ^{18}O due to their more maritime locations in the Western Eurasian Arctic and therefore milder climate conditions compared to SZ. Some of the slight deviations between AN and Austfonna δ^{18}O records might be caused by the lower sampling resolution (about 25 cm) of the Austfonna ice core and its dating by a Nye model not taking into account
25 the probable growth of the ice cap analogously to AN ice cap. The strong accordance of both AN and Austfonna δ^{18}O records to that of Vetreniy ice cap, however, decreases before 1800 and is lost before about 1700. As FJL is situated between Svalbard and SZ the Vetreniy ice core should reflect the same major SAT patterns also before 1700. We assume that below the Laki reference horizon (1783) the Vetreniy ice core age model

progressively overestimates the age by up to about 100 yr in the oldest part of the record. A possible offset between different dating approaches for this core is already noticed by Henderson (2002). The δ^{18}O record from the high-altitude Lomonosovfonna (1250 m a.s.l.), the best-studied Svalbard ice core (e.g. Isaksson et al., 2005; Divine et al., 2011) exhibits, conversely, almost no accordance to our AN record, and receives most probably different atmospheric signals due to the higher altitude level compared to the other three ice-core locations. However, with our new AN δ^{18}O record we are able to expand the SAT information for the Barents and Kara seas region back to 900.

Even though our AN δ^{18}O record displays the same overall trends as the Arctic-wide reconstruction of Kaufman et al. (2009) there are beside some accordance several differences between both records, in particular regarding the timing and peculiarities of several specific events. For instance the SAT minimum around 1800 as well as the following ECTW occurred distinctly earlier by some decades and more pronounced in the AN δ^{18}O record as compared to the Arctic scale (Fig. 5). There seem to be mainly two possible explanations for these differences: spatial and/or seasonal effects.

As the reconstruction of Kaufman et al. (2009) is based predominantly on records from the North American Arctic and Greenland it is likely spatially weighted and its significance is lower for the Eurasian Arctic. This again shows the particular climate regime of the Barents and Kara seas region framed by the AN and Austfonna ice-core sites as well as distinctly different SAT patterns in the Eurasian and North American Arctic. The lead of the Western Eurasian Arctic SAT might indicate that such major SAT changes could have originated in the Barents and Kara seas region and progressively expanded to the east, reaching the North American Arctic only some decades later. This is supported by the later ETCW occurrence in the North American Arctic also in meteorological data (Polyakov et al., 2003b).

On the other hand, the reconstruction of Kaufman et al. (2009) represents mainly summer temperatures and the AN δ^{18}O record annual temperatures. So, also seasonal effects might contribute to these differences. This assumption is supported by a recent study of Sidorova et al. (2013) that shows also a lead of AN δ^{18}O over tree-ring proxies

from the nearby Taimyr Peninsula as well as Northern Yakutia for the 1800 SAT minimum and the subsequent warming to the ETCW. Moreover, in particular the first part of ETCW occurred predominantly in winter (Polyakov et al., 2003b) and was consequently recorded in our annual AN δ^{18}O record but not in the summer SAT proxies used by Kaufman et al. (2009) and Sidorova et al. (2013).

Although our AN δ^{18}O data indicate higher temperatures around 900 and in the 13th century than in the 17th century, no distinct long-lasting climate epochs such as the Medieval Climate Anomaly or the Little Ice Age (LIA) are readily detectable in the AN δ^{18}O record (Fig. 5). Only the cold period around 1800 and the subsequent warming could be interpreted as final stage and termination of the LIA. As the same is true for the Austfonna δ^{18}O record, we conclude that the Barents and Kara seas region did not experience a pronounced LIA. In contrast, the AN δ^{18}O record exhibits several abrupt cooling and warming events such as in the 15th and 16th as well as in the 18th and 20th centuries (Fig. 5). These events exceed the dominant range observed in our record and are of regional significance as they are also detectable in the Austfonna and partly in the Vetreniy ice cap δ^{18}O records. The range and change rates in our AN δ^{18}O time series are comparable to the SAT rise before 1920 leading to the ETCW.

These abrupt changes and resulting cool and warm periods might also be triggered by internal climate variability as assumed for the ETCW, i.e. related to changes in the prevailing wind patterns caused by shifts between dominant atmospheric circulation types as well as to corresponding sea-ice feedbacks in the Barents and Kara seas region. This is supported by according changes in the AN sodium record (Fig. 5) as well as by model results for the late 15th century (Crespin et al., 2009). However, the existing sea-ice records are either too short as for the Barents Sea (Vinje, 1999) or show only slight variations but no according patterns for the 15th and 16th centuries (Fig. 5), probably because they represent Arctic scale (Kinnard et al., 2011). However, the period of abrupt changes in the AN and Austfonna δ^{18}O records (Fig. 5) fits well to a phase of significantly increased strengths of Icelandic Low and Siberian High from the 15th to the 20th century (Meeker and Mayewski, 2002). As both centers of action

dominate the atmospheric circulation from the North Atlantic to Asia, their joint higher variability is assumed to lead to rapidly changing patterns of air-mass advection and consequently for the occurrence of such abrupt SAT changes in the Western Eurasian Arctic.

Whereas our AN δ^{18}O and the Arctic-wide SAT (Kaufman et al., 2009) records display a marked similarity in the 11th and 12th century, they show contrary trends during the 15th and 16th centuries (Fig. 5). Abrupt cooling events in the Barents and Kara seas region are accompanied by warming events in the Arctic scale and vice versa. The causes for these differences might be comparable to those already discussed for the periods around 1800 and in the 20th century. This pattern could be interpreted as a kind of SAT see-saw on spatial (Eurasian vs. North American Arctic) and/or seasonal (annual vs. summer) scale.

5 Conclusions

The results presented in this paper highlight the potential of the AN ice core as high-resolution climate archive for the Late Holocene, i.e. the last about three millennia. Beside a long-term decrease due to ice-cap growth and climate cooling the AN δ^{18}O record shows evidence for major temperature changes over the last millennium that are representative at least for the Western Eurasian Arctic, i.e. the Barents and Kara seas region. Of particular importance are several abrupt warming and cooling events leading e.g. to the absolute SAT minimum around 1800 and the absolute SAT maximum in the early 20th century. The latter exhibits a specific double-peaked shape typical for the Barents and Kara seas region. These abrupt changes might be caused by internal climate dynamics related to shifts of atmospheric circulation patterns and according sea-ice feedbacks.

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Eurasian Arctic climate over the past millennium

T. Opel et al.

Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

|◄ | ►|

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Grant, A. N., Brönnimann, S., Ewen, T., Griesser, T., and Stickler, A.: The early twentieth century warm period in the European Arctic, Meteorol. Z., 18, 425–432, doi:10.1127/0941-2948/2009/0391, 2009.

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Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

|◀ | ▶|

◀ | ▶

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Meyer, H., Schönicke, L., Wand, U., Hubberten, H. W., and Friedrichsen, H.: Isotope studies of hydrogen and oxygen in ground ice – experiences with the equilibration technique, Isot. Environ. Healt. S., 36, 133–149, doi:10.1080/10256010008032939, 2000.

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Eurasian Arctic climate over the past millennium

T. Opel et al.

Title Page

Abstract	Introduction
Conclusions	References
Tables	Figures

|◀ | ▶|
◀ | ▶

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

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Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

|◄ | ►|

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Table 1. Correlation coefficients (annual data and 5 yrm values) between AN δ^{18}O and meteorological time series. Displayed are the values for detrended time series and in brackets for the original records. Bold (italic) numbers indicate correlation coefficients statistically significant at the level of $p = 0.05$ ($p = 0.1$), verified by an one-tailed t test taking into account the reduced degrees of freedom due to autocorrelation (lag 1).

time series	Time period	r_{annual}	r_{5yrm}
Vardø annual	1840–1998	**0.20 (0.38)**	**0.55 (0.76)**
Vardø MAM	1840–1998	0.11 **(0.30)**	0.20 **(0.59)**
Vardø JJA	1840–1998	*0.13* **(0.16)**	*0.43 (0.43)*
Vardø SON	1840–1998	**0.15 (0.30)**	**0.51 (0.69)**
Vardø DJF	1841–1998	**0.19 (0.34)**	**0.46 (0.69)**
Arkhangelsk annual	1814–1998[a]	**0.19 (0.27)**	**0.51 (0.57)**
T_{NA} annual	1802–1998[b]	**0.20 (0.54)**	*0.47* (0.78)
Arctic annual	1875–1998	**0.43 (0.55)**	*0.72* (0.81)

[a] Missing data: 1832–1833 and corresponding 5 yrm values.
[b] Missing data: 1803–1806, 1822–1823, 1825, 1828 and corresponding 5 yrm values.

Eurasian Arctic climate over the past millennium

T. Opel et al.
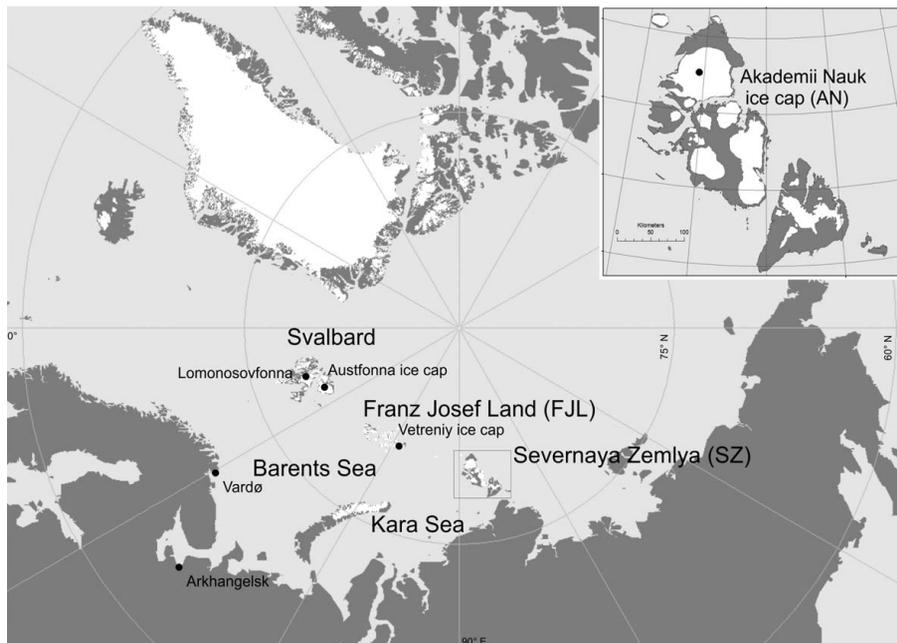
Fig. 1. Map of the Eurasian Arctic (inset: Severnaya Zemlya with the drilling point at Akademii Nauk ice cap). Locations mentioned in the text are labeled.
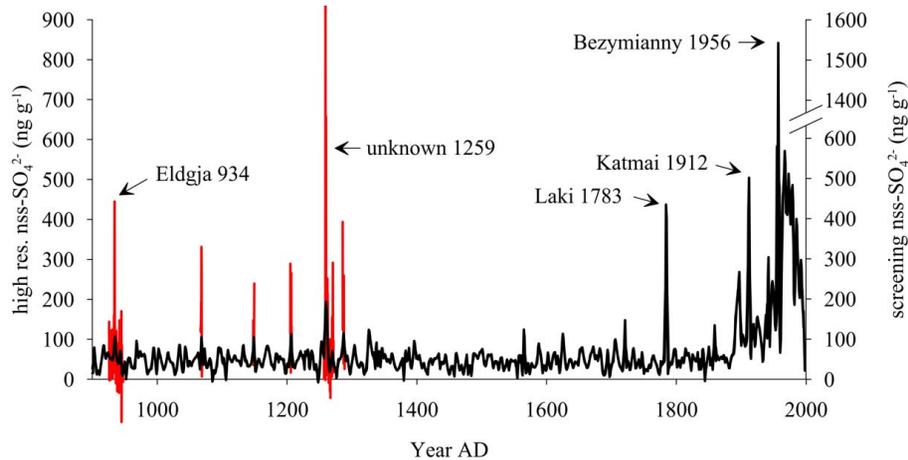
Fig. 2. AN nss-SO$_4^{2-}$ record with annual values based on screening data in black (right axis), high resolution data for selected core sections in red (left axis). Peaks interpreted as volcanic reference horizons are labeled.
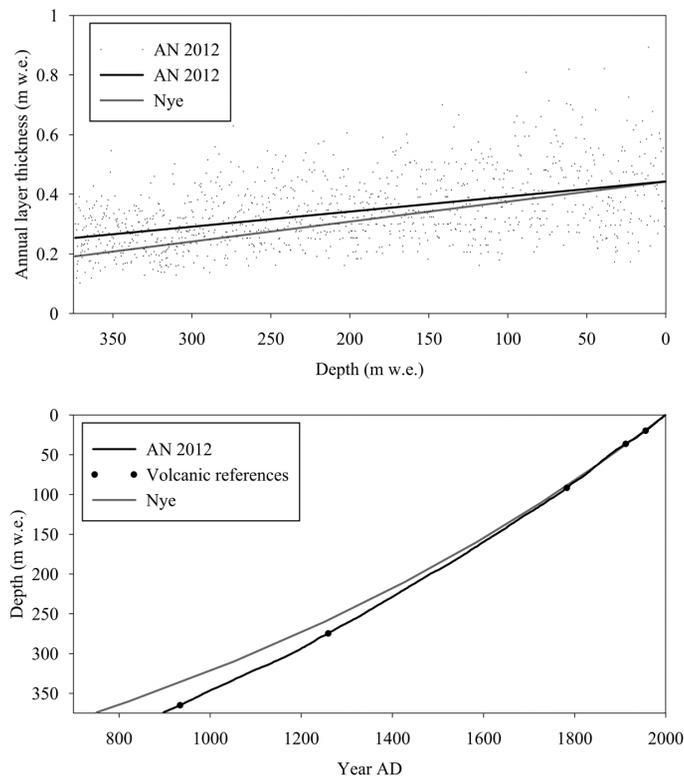
Fig. 3. (Top) Annual-layer thickness of AN ice core (dots) derived from stable-isotope based layer counting, linear fit of annual-layer thickness (black line) and annual-layer thickness derived from a Nye model (grey line). (Bottom) Depth-age relations of core chronology AN 2012 (black line) and based on a Nye model calculated with an accumulation rate of 0.44 m w.e. and an ice-cap thickness of 660 m w.e. (grey line) for the studied core section. Black dots represent volcanic reference horizons.
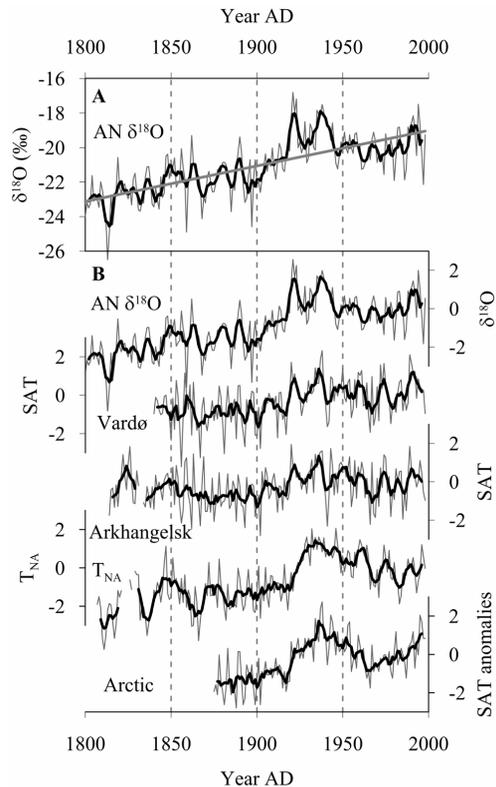
Fig. 4. AN δ^{18}O record compared to meteorological data and compilations. **(A)**: original AN δ^{18}O record (including linear trend in grey). **(B)** (From top to bottom): Standardized time series of AN δ^{18}O, Vardø and Arkhangelsk SAT (Brohan et al., 2006), Atlantic-Arctic boundary region SAT anomalies (T_{NA}) (Wood et al., 2010) and Arctic SAT anomalies (Polyakov et al., 2003b). For better comparison the time series are plotted as normalized deviations (z) from the 1900–1998 mean (m): $z = (x - m)s^{-1}$, where s is the standard deviation of the raw series for the period 1900–1998. Displayed are annual values (grey line) and 5 yrm values (thick black line).

Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

|◄ | ►|

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion
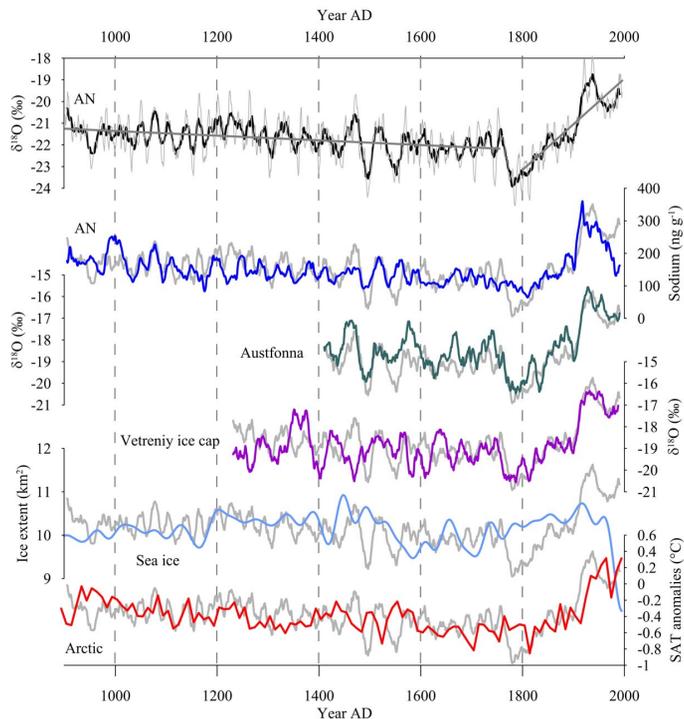
Fig. 5. AN δ^{18}O (including linear trends for 900–1760 and 1800–1998) compared to (from top to bottom) AN sodium concentrations, Austfonna δ^{18}O (Isaksson et al., 2005), Vetreniy ice cap δ^{18}O (Henderson, 2002; data from Kinnard et al., 2011), Arctic sea ice extent (Kinnard et al., 2011) and Arctic SAT anomalies (Kaufman et al., 2009). Displayed are 5 yrm values for AN δ^{18}O (thin line), 15 yrm values for AN δ^{18}O, AN sodium, Austfonna δ^{18}O, Vetreniy ice cap δ^{18}O, Arctic sea ice (thick lines) as well as 10 yr means for Arctic SAT (thick line). For easier comparison, to each graph the AN δ^{18}O 15 yrm record is added in light grey in the same scale as above and adjusted for the best fit of the 20th century maximum (except for sea ice).

Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

|◄ | ►|

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion