POTENTIAL TO RECOVER CLIMATIC INFORMATION FROM SCANDINAVIAN ICE CORES: AN EXAMPLE FROM THE SMALL ICE CAP RIUKOJIETNA

BY
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ABSTRACT. We have studied a 33.7 m deep ice core from a small polythermal Scandinavian ice cap to determine whether it is possible to recover pre-20th century climatic information from the glacier. Ice structural studies show a significant change from clear ice above 11 m depth (superimposed ice indicating refreezing) to bubbly ice below 11 m depth, indicating this is the transition between Little Ice Age (LIA) and 20th century ice. Calculations with a Nye-age model, along with a mass balance reconstruction, show that this structural boundary likely formed in the last part of the LIA, which in this region ended about 1910. The ice below this boundary was sampled and analysed for stable isotopic composition and ionic content, which both show significant variations with depth. The stable isotope record likely contains cycles of annual duration during the LIA. The chemistry in the ice core indicates that the information is useful, and can be used to interpret climatic and environmental variables during the LIA. A comparison of Riukojietna ion chemistry and oxygen isotope records with similar records from other glaciers in this region reveals a clear continental–maritime gradient. Changes in this gradient with time may be possible to resolve using such ice core records. Results from this study demonstrate that ice cores from glaciers in this climatic environment can be useful in revealing environmental conditions from climatically colder periods and yield pre-industrial benchmark values for chemical loading and oxygen isotopes, but that hiatuses complicate the depth–age relationship.

Introduction

Glaciers located in sub-polar and temperate regions contain potential archives of regional environmental and climate change. These archives can be sampled through ice coring. The main reason why such glacier archives have remained relatively unexplored is that today they are subjected to a warmer climate than the polar or high elevation ice sheets and ice caps normally investigated. The closer the average snow and ice temperature is to the melting point, the larger is the risk for alteration or destruction of the physical and chemical signals within the ice.

Studies in the 1970s of ice cores from wet and relatively warm regions such as Iceland and Mt Kenya showed that ice cores from these regions lack detectable climatic and environmental signals (Arnason 1981; Thompson 1981). This is because meltwater percolates into the snow pack if the accumulation area is in the percolation zone or wet snow zone (Paterson 1994). Under these conditions, the original chemical and isotopic variations related to climate and environmental variability become severely disturbed. It might be noted that water isotope stratigraphy appears more robust than the chemical record (Pohjola *et al.* 2002b) and therefore has a higher potential to be useful as a climatic proxy (e.g. Oerter *et al.* 1985; Naftz *et al.* 1996).

Other studies at low altitude polar and sub-polar ice fields (e.g. Koerner and Fisher 1990; Isaksson *et al.* 2001) and on high altitude ice fields in lower latitudes (e.g. Thompson *et al.* 1995; Schwikowski *et al.* 1999) have pushed the limit at which valuable ice core records may be extracted from the dry-snow zone to the percolation zone of a given ice mass. The study by Holmlund (1998), who re-

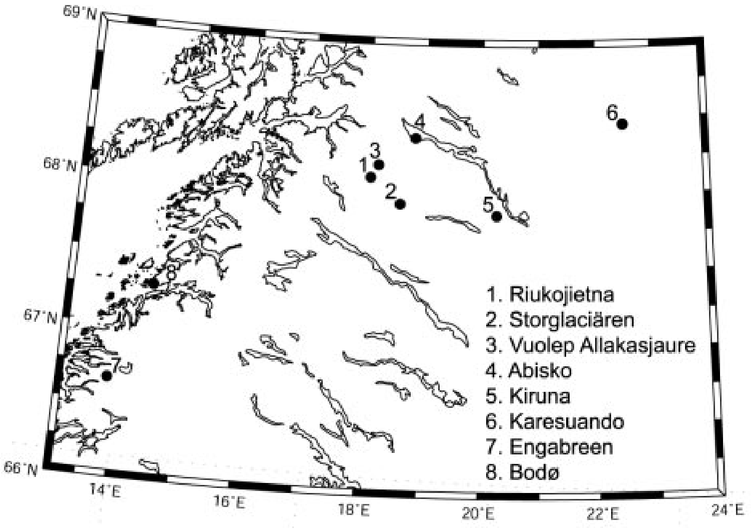

Fig. 1. Location of places mentioned in the text.

Fig. 2. An oblique aerial photograph of the ice cap Riukojietna (photo by P. Jansson 2003). The drill site is marked with an arrow.

trieved an ice core from the valley glacier Mårmaglaciären in 1997, illustrates the problems involved in interpreting ice cores in Scandinavia.

There are many small ice caps on the Scandinavian Peninsula which can potentially offer records of climatic and environmental proxies for this region. A 33.7 m ice core from the small ice cap Riukojietna retrieved in April 1996 from an elevation of 1450 m a.s.l., provides an opportunity to assess this potential. Because we expected the record to be punctuated by hiatuses arising from periods of negative mass balance, we applied a non-systematic sampling strategy to assess the locations of the 'best' ice core sections. We then used all available data to gauge the potential of these sections for regional palaeoclimate reconstruction, i.e. through an analysis of the ice structure and chemistry. Using these and mass balance considerations, we arrived at the conclusions of the presence of **Little Ice Age** (**LIA**) climate variability and a potential for such ice cores to yield pre-industrial benchmark values for chemical loading and oxygen isotopes.

Geographical and physical setting

Geography and Holocene history

The Riukojietna ice cap is situated in the northern part of the Scandinavian mountain range on the border between Norway and Sweden, approximately 20 km west of the main topographic divide

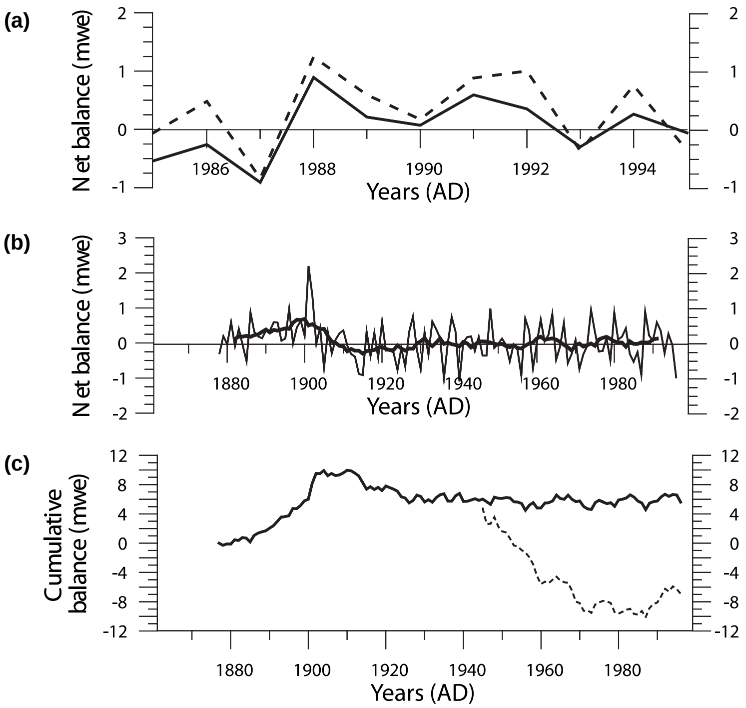

Fig. 3. (a). Measured net balance on Riukojietna (solid line) and Storglaciären (dashed line) for 1985–1995. (b). Reconstructed net balance of Riukojietna 1878–1996 using the Karesuando July temperature and the Bodø February precipitation (Fig. 1). The thin line is the reconstructed annual net balance; the thicker line is a nine-point running mean. (c). The thicker line is the cumulative Riukojietna net balance calculated from 1878. The dashed line is the measured cumulative net balance of Storglaciären, where the start point is set in level with the Riukojietna balance.

and 100 km east of the coast of Norway (Figs 1, 2). The ice covers a granite / hard schist plateau at 1300 m a.s.l. Riukojietna measures approximately 4.6 km^2, has a maximum thickness of 105 m (Holmlund *et al.* 1996), an estimated volume of 0.17 km^3, and extends from 1140 to 1456 m a.s.l. (Rosqvist and Østrem 1989). We chose to core this glacier on the basis of its geometrical simplicity (because the domed shape of the glacier suggests minimal disturbance of ice layers due to ice flow) and logistical convenience. The ice core was recovered near the summit, where the ice thickness measured 80 m (Tarfala 2004). Our 1986 thermistor measurements indicated that the base of the ice cap was frozen to the bed at the coring site.

Similar to many other glaciers in Scandinavia, Riukojietna was small and inactive, or, perhaps more likely, non-existent, during the early Holocene climatic optimum (8000–6000 calendar years before present) (Karlén 1981). However, lacustrine sediments derived from melt water streams off Riukojietna indicated that the ice cap was reactivated 2500 years ago and has been in existence since. The presence of a terminal moraine indicates that at some point Riukojietna advanced to a Holocene maximum position. Karlén (pers. comm.) used lichenometry to infer a date of *c*. 1910 for this terminal moraine, an age which is similar to other, better dated, LIA moraines in this region (Karlén and Denton 1976). The eastward-draining glacier tongue that deposited the terminal moraine subsequently receded 350 m until 1963 (Schytt *et al*. 1963), and another 450 m between 1960 and 1978 (Rosqvist and Østrem 1989).

A comparison of two topographic maps of Riukojietna (Rosqvist and Østrem 1989) revealed that the summit elevation had lowered 9 ± 2 m between 1960 and 1978, probably due to sustained negative mass balances. Today, the highest net balance usually occurs around 1400 m a.s.l. on the eastern flank of the summit (lee-side effect).

Mass balance trends during the last century

In the previous section we examined the geological evidence for former extensions of Riukojietna. In this section we use mass balance information from Riukojietna and nearby Storglaciären to provide links between stratigraphy and time.

Mass balance measurements on Riukojietna were initiated in 1986 (Fig. 3a; Rosqvist and Østrem 1989), and the record is available from Tarfala (2004). The most extensively studied glacier in this region is Storglaciären, situated 50 km southeast of Riukojietna (Fig. 1), with a mass balance record extending back to 1945 (Holmlund *et al.* 1996). Storglaciären has a similar size to Riukojietna (3.1 km^2), but spans a larger elevation range (1130–1700 m a.s.l.). The annual net balance on Riukojietna was on average 0.38 m **water equivalent** (**m w.e.**) lower than on Storglaciären between 1985 and 1995, while the trend was similar (Fig. 3a).

From the Storglaciären mass balance record we know that it has predominantly experienced negative net balances between 1945 and 1987, after which the glacier gained mass again (Fig. 3c; Holmlund 1998; Tarfala 2004). Both Storglaciären and Riukojietna have had stable frontal positions between 1989 and 1994 (Holmlund *et al.* 1996), which further strengthens the similarities between the two glaciers.

Earlier reconstructions of mass balance in this region are based on maps and photographs, and allow the following generalized picture to be presented. Local glacial maxima were attained around 1910, at the culmination of the last LIA advance (Svenonius 1910; Holmlund 1987), followed by a retreat of glacier margins until 1989. Hence, these observations are in tune with inferred predominantly positive net balances until 1910, negative net balances until 1987, and, finally, positive net balances to force the retreating glacier margin to a halt. We attempt to quantify this mass balance trend for Riukojietna using a multiple regression analysis of the 12-year net balance record (Fig. 3a) against climatic data from various stations in northern Scandinavia. The best results were achieved using the average July temperature at Karesuando and February precipitation at Bodø ($r^2 = 0.80$), allowing a reconstruction of the mass balance back to 1878 (Figs 1, 3b). Our reconstruction is correlated to the measured Storglaciären mass balance with $r^2 = 0.40$ (1945–1996), which is a relatively poor match. A comparison of the 1945–1996 cumulative mass balance records of Storglaciären and Riukojietna (Fig. 3c) reveals that our reconstruction is deficient in yielding a reliable long-term trend of the mass balance, despite the relatively good correlation between the climatic parameters and mass balance for the last 12 years.

We conclude that Riukojietna must have accumulated ice for an unknown period of length up to *c*. 1910. A long period of overall negative net balances reduced the volume and length of Riukojietna subsequently, although there were likely some years when superimposed ice accumulated in the upper parts of the glacier. Riukojietna has experienced mostly a positive net balance between 1988 and the drilling of the core in 1996.

Methods

The ice core was retrieved using a thermal core drill with a 2-m long barrel having a diameter of 10 cm, and an effective drilling speed of 1–2 m h^{-1}. Heated core dogs were used to snap off the ice core before the core was hauled up. The recovered core is 33.7 m long and core loss was 2.4 m, of which the largest part occurred between 5.0 and 6.9 m depth. The core sections were wrapped in plastic bags, stored in a snow cave during fieldwork, and subsequently transported to a freezer and laboratory at the Tarfala Research Station (1 km from Storglaciären, Fig. 1) where the ice core structure was described and the ice core was sub-sampled.

Quantitative structural description of the ice follows Pohjola *et al.* (2002b) and was performed on the flat cut ice core surface with up to 5 mm resolution. Density was measured on the snow and firn section of the core, using the standard gravimetric method. The relative ice density profile further down was estimated using the ice facies following procedures outlined by Pohjola *et al.* (2002b): 910 kg m^{-3} for *clear ice* (ice devoid of air bubbles), 895 kg m^{-3} for *diffuse ice* (ice with few, mostly large, bubbles which are often arranged in arrays), and 850 kg m^{-3} for *bubbly ice* (ice with many bubbles). The structural differences are indicative of the degree of melting of the firn that formed these ice facies.

The upper 14.3 m of the core was sub-sampled at 5–15 cm increments following structural boundaries; between 14.3 and 15.4 m and between 24.0 and 26.0 m the core was sub-sampled at 4–6 cm. Of these, a total of 136 samples, representing all ice facies, were analysed for oxygen isotope composition ($\delta^{18}O$) on a Finnegan Delta stable isotope mass spectrometer at the Byrd Polar Research Center of Ohio State University.

Two core sections (14.3–15.4 m and 24.0–26.0 m) were cut into 54 samples and were measured by ion chromatography at the Byrd Polar Research Center for concentrations of general anions and cations. The procedures for ice core decontamination and ion analysis follow Cole-Dai *et al.* (1995).

Liquid conductivity of melted samples was

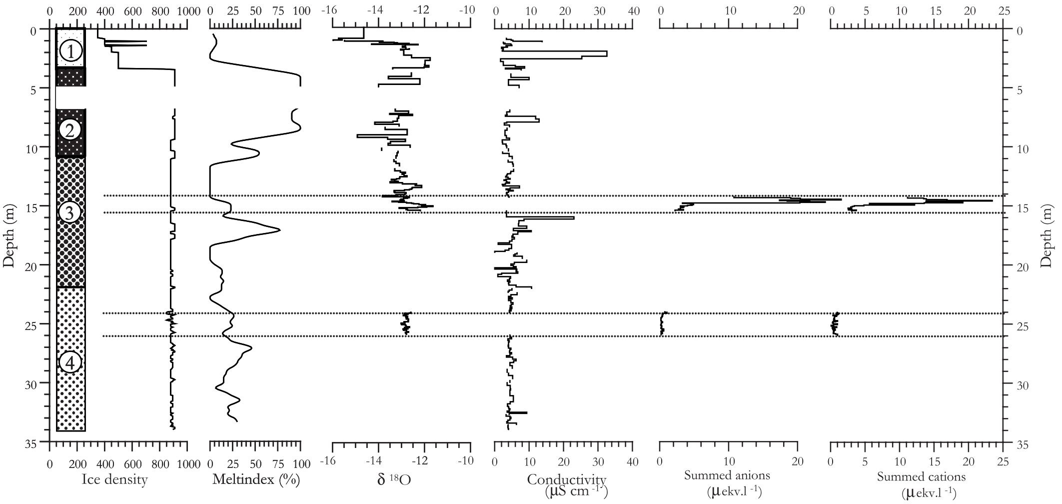

Fig. 4. Log of the Riukojietna ice core showing structure, ice density, melt index (11-point running mean), $\delta^{18}O$, liquid conductivity, and major anions and cations. The anions are the summed concentrations of Cl^-, NO_3^- and SO_4^{2-} and the cations are the summed concentrations of Na^+, K^+, NH_4^{2+} and Ca^{2+}. Sections lost during drilling are left white in the core log. Dotted lines mark the sections sampled for ion chemistry and hence those missing in the conductivity record.

measured with a self-calibrating probe. Before measurement, the temperature of the melted sample was equilibrated to room temperature.

Thin sections were prepared of five levels in the lower part of the core (27.2, 29.7, 30.5, 32.7, and 33.6 m) for texture analysis and ice crystal size estimates.

Results and interpretation

Stratigraphy and texture

Based on the ice structural analysis, the core can be divided into four main sections (Fig. 4; Table 1). Section 1 includes the upper 3.4 m (or 1.5 m w.e.) of snow and firn, and a few 2–3 cm ice lenses. Field observations indicated that the uppermost 1.1 m (0.3 m w.e.) represents the 1995–1996 winter accumulation. Section 2 (3.4–11.0 m; 6.9 mwe) is almost entirely composed of a *clear ice* facies, section 3 (11.0–21.7 m; 9.5 m w.e.) consists of *bubbly ice* interspersed with some *clear ice*, and section 4 (21.7–33.7 m; 10.8 mwe) is dominated by *bubbly ice* interlaced by thin layers of *diffuse* and *clear ice*. Overall, about 50% of the core consists of the *bubbly ice* facies.

The structural description can be used to construct a melt-index (Koerner and Fisher 1990). This index depicts the ratio between ice affected by melting and ice unaffected by melting. A high melt index indicates that melting and refreezing proc-

Table 1. Minimum, average, and maximum values of $\delta^{18}O$, melt index, and conductivity in the four structurally different sections of the Riukojietna ice core.

Section	Depth interval (m)	$\delta^{18}O$ Min.	Av.	Max.	Max.–Min.	Melt index (%)	Conductivity (µS cm^{-1})
1	0–3.4	–16.0	–13.4	–11.8	4.2	3	10.9
2	3.4–11.0	–14.9	–13.0	–12.2	2.7	84	4.9
3	11.0–21.7	–13.8[1]	–12.9[1]	–11.6[1]	2.2[1]	14	4.9[3]
4	21.7–33.7	–13.0[2]	–12.8[2]	–12.6[2]	0.4[2]	22	4.6[3]
1–4[4]	0 – 33.7		–13.0			32	5.0

[1] Interval covers 11.0–15.4 m.

[2] Interval covers 24.0–26.0 m.

[3] Interval excluding parts sampled for chemistry (cf. Table 3).

[4] Values here are weighted averages for the whole core.

esses affected a high percentage of the accumulated ice. The highest melt index occurs between 3.4 and 11.0 m (Fig. 4; Table 1). Below 11 m the melt index is generally lower except for an interval between 16 and 18 m.

From the higher melt index in the upper part we infer that the upper third of the core suffered more melt and percolating water than the lower two-thirds of the core. The upper third of the core contains two-thirds of the *clear ice* facies. This suggests that melting, infiltration and refreezing were more frequent and vigorous in the time period represented by section 2.

Ice textural analysis on the lowest part of the core revealed the occurrence of ice crystals with areas of ≥5 cm^2. Although we do not know at what rate ice crystals grow within Riukojietna, the size of the crystals indicate that these grains grew over a long time period or that they experienced large strains.

Oxygen isotopes

The average isotopic composition (δ^{18}O) for the ice core is –13.0‰. The average δ^{18}O increases with depth (Table 1), i.e. more enriched δ^{18}O occurs in the lower sections. The δ^{18}O variability, which on average is 0.8‰, is highest (*c.* 2%) in the upper part of the core and very low, e.g. 0.25‰ at 24.0 m, in the lowest part of the core (Fig. 4). The decreasing δ^{18}O amplitude with depth is expected due to the vapour-driven diffusion of water molecules (Johnsen *et al.* 2000), or perhaps homogenization due to percolation (Arnason 1969).

The most negative δ^{18}O values are found in the upper 11 m of the ice core. Normally, a change towards more negative δ^{18}O values is interpreted to be a result of a lowering of atmospheric temperatures. However, such a change is not consistent with available temperature records from this region which show a slight warming during the last decades (Klingbjer and Moberg 2003). We find it most likely that variations in snowmelt and its effect on the relative contribution of summer and winter precipitation in the record has caused the observed δ^{18}O distribution. This implies that ice in sections 3 and 4, which are characterized by low-melt indexes, probably contains a higher proportion of isotopically heavier summer snowfall. More of the annual signal is therefore retained here in contrast to sections 1 and 2 that likely have lost most of the summer precipitation by run-off, and therefore contain a 'colder' signal.

Table 2. Available δ^{18}O records from northern Sweden and Norway.

δ^{18}O site	Annual average	Annual min.	Annual max.	First quartile	Fourth quartile	Seas. ampl.
Abisko[1]	–13.2	–17.1[6]	–10.6[8]	–14.4	–11.2	1.6[10]
Kiruna[1]	–14.1	–17.8[6]	–10.9[8]	–17.2	–10.7	3.2[10]
Riukojietna	–13.0[5]	–16.0[7]	–11.8[9]			2.1[11]
Storglaciären[2]	–13.8[5]					
Engabreen[3]	–8.7[5]					
Vuolep Allakasjaure[4]	–13.9					

[1] This data is based on monthly averages 1975–1980 from the GNIP database (IAEA/WMO, 1998).

[2] Averaged over 56 samples, including surface samples (Glasser et al. 2003) and ice core samples from the lowest ablation area (Pohjola et al. 1994).

[3] Averaged over 96 supra-, en- and subglacial samples (Jansson et al. 1996).

[4] Average of 12 samples from a lake located *c.* 10 km north of the ice cap (Rosqvist et al. 2004).

[5] Annual average is here assumed to be equal to the average of all samples in the ice core.

[6] December average of the last 6 years.

[7] Minimum value of last year's snow pack.

[8] June average of the last 6 years.

[9] Maximum value of last year's snow pack.

[10] Determined as difference between first and fourth quartiles.

[11] Determined from the maximum and minimum data.

The δ^{18}O value of –13.4‰ for the uppermost 3.4 m of snow and firn (section 1) in the Riukojietna ice core is similar to the modern isotopic composition in Abisko precipitation (–13.2‰) (*c.* 50 km northeast of Riukojietna; Fig. 1) and to the contemporary isotopic values obtained on through-flow water in the area (Table 2) (IAEA/WMO 1998; Shemesh *et al.* 2001; Hammarlund *et al.* 2002; Rosqvist *et al.* 2004).

In contrast to the records discussed above, the isotopic composition of precipitation from Kiruna is more depleted (Table 2). This can be explained by the fact that Kiruna is located 100 km southeast of Riukojietna, leeward of the mountains, and thus receives more depleted precipitation from air masses originating from the North Atlantic due to rain-out effects. Kiruna also receives proportionally more δ^{18}O depleted precipitation derived from air masses from the east and southeast of Riukojietna, which is mainly fed by westerlies. This east-west gradient also appears in comparison to the isotopic composition of ice sampled from Storglaciären and Engabreen, the latter located *c.* 230 km SW of Riukojietna (Fig. 1). Ice from Engabreen only has an average δ^{18}O value of –8.7‰, while Storglaciären ice has a mean δ^{18}O value of –13.8‰ (Table 2). The

Table 3. Average concentration of anions, cations, and ion indices for two different sections of the Riukojietna ice core, and averages for the two chemically different parts in the upper section. The anions Cl^-, NO_3, SO_4^{2-} and the cations Na^+, K^+, NH_4^+, Mg^{2+}, Ca^{2+} were measured, while the ratios were calculated. All values are in μeq. l^{-1}.

Depth interval (m)	n	Cl^-	NO_3	SO_4^{2-}	Na^+	K^+	NH_4^+	Mg^{2+}	Ca^{2+}	Mg^{2+}/Na^+	SO_4^{2-}/Cl^-	SO_4^{2-}/NH_4^+	Na^+/Cl^-
14.0–14.8	12	16.02	1.05	1.11	9.59	0.97	2.33	0.58	1.02	0.06	0.07	0.48	0.60
14.8–15.4	9	2.63	0.14	0.76	2.71	0.22	0.68	0.34	0.38	0.14	0.29	1.11	1.03
14.0–15.4	21	8.72	0.55	0.92	5.84	0.56	1.43	0.45	0.67	0.10	0.11	0.64	0.67
24.0–26.0	33	0.25	0.13	0.05	0.05	0.11	0.02	0.39	0.02	10.97	0.20	2.48	0.20
0–34	54	3.58	0.29	0.39	2.32	0.29	0.57	0.41	0.28	6.09	0.11	0.68	0.64
EB[1]	2				16.5	11.1		1.15	5.5				
SG[2]	15	1.2	1.6	1.8	1.7	0.8							1.4

[1] These samples were taken from surficially exposed englacial ice off the Engabreen ice front (Jansson *et al.* 1996).

[2] These samples were taken from shallow ice cores on the frontal part of the ablation area of Storglaciären (Pohjola *et al.* 1994)

mean annual atmospheric temperatures over the glaciers are not known but an estimate using annual mean atmospheric temperatures at the Tarfala Research Station (Storglaciären) and at Bodø (Engabreen) (using an average lapse rate of 0.7°C / 100 m) yields that the accumulation area of Storglaciären is at least 1°C cooler than the accumulation area of Engabreen. Keeping in mind that the relationship between δ^{18}O and temperature for this temperature interval often equals *c*. 1‰ per °C^{-1} (Souchez and Lorrain 1991), this indicates that the proximity of Engabreen to the Norwegian Sea explains a large part of the difference in δ^{18}O between these places.

However, because the ice samples were collected on the tongues of Storglaciären and Engabreen, they could be several hundred years old (e.g. Pohjola 1996) and the isotopic difference could therefore reflect the fact that we compare climate conditions at different times.

Chemical information

The two sections analysed for chemical composition were chosen because the stratigraphical analysis indicated that these sections experienced minimal post-depositional, melt-related, alteration (Fig. 4). Earlier studies on Scandinavian glaciers have shown that the ice chemistry is much affected by melting (Raben and Theakstone 1994). Our assumption is that ice from earlier and colder climate periods retains a less altered environmental signal. Among the anions, average Cl^- concentrations are significantly higher than either average NO_3^- or SO_4^{2-} concentrations (Table 3). Likewise, average concentrations of Na^+ and NH_4^+ are significantly higher than those of the other cations. Total cation concentrations are higher than total anion concentrations for most of the samples (Table 3), requiring the presence of other anions, most likely HCO_3^-. This is because HCO_3^- is the only inorganic acid not accounted for by the ion chromatography, and is the most likely candidate to maintain ionic balance.

The two sections differ in ionic loading. The upper section (14.0–15.4 m) show relatively high loadings, and the lower section (24.0–26.0 m) shows very low concentrations. The two sections are similar with respect to the melt index; both sections were taken in areas of little sign of infiltrated meltwater (i.e. clear ice). Could the difference in chemical loading still be explained by different action of melting between these sections? Even if no percolated water is detected, the upper section could have lost more water than the lower section due to meltwater run-off. Iizuka *et al.* (2002) suggested that ion ratios can be used to demonstrate effects of percolation in ice core studies. Specifically, they found that because Mg^{2+} and Na^+ are derived from sea salts (at least in maritime environments), their concentrations should maintain a sea salt ratio of $Mg^{2+}/Na^+ = 0.24$ μeq l^{-1} (0.12 μmol l^{-1}), or in snow and ice in the absence of melting. Since Mg^{2+} is more mobile (i.e. elutes more readily) than Na^+, it follows that ice which has experienced melt and percolation will show Mg^{2+}/Na^+ values lower than 0.24. Indeed, in the Riukojietna ice the Mg^{2+}/Na^+ ratio is only 0.10 in the 14.0–15.4 m interval, while ice at 24.0–26.0 m reveals a ratio of 10.97 (Table 3). This suggests loss by melt from the upper section, but argues against loss of ions due to melt in the 24.0–26.0 m section. Inspecting the upper section we find that this section can be divided between 14.0–14.8 m having high chemical loadings, and

14.8–15.4 m having low concentrations (Fig. 4, Table 3). We find that the melt index is lower in the upper part of the section, and higher below, suggesting that some water has infiltrated the lower part. The low Mg^{2+}/Na^{+} in the 14.0–14.8 m part, and the higher values in the 14.8–15.4 m part are in accordance with such an explanation. Another way to explain deviation from sea salt ratio in the Mg^{2+}/Na^{+} ratio is import of continental ions into the ice, which may be a more important contribution at Riukojietna than on Svalbard, the environment referred to by Iizuka *et al*. (2002).

We also test for the degree of percolation and infiltration in the studied ice core samples using the following ratios, SO_4^{2-}/Cl^{-} and SO_4^{2-}/NH_4^{+}. Iizuka *et al*. (2002) suggested SO_4^{2-}/Cl^{-} should yield similar results to the Mg^{2+}/Na^{+} ratio, except that SO_4^{2-} normally also is influenced by other sources than the ocean. Pohjola *et al*. (2002b) showed that SO_4^{2-} in a Svalbard ice core elutes more readily than most of the major ions, while NH_4^{+} was the most conservative ion. This is because the majority of NH_4^{+} is situated in the ice crystal matrix, while SO_4^{2-} is preferably pushed out to the ice crystal boundaries during crystal growth. These two additional ratios depict the same picture of significant loss of mobile ions at the 14.0–14.8 m level, less or no loss at the 14.8–15.4 m level, and enrichment of the mobile ions at the 24.0–26.0 m level (Table 3).

To complicate the analysis above we know that the total chemical loading is higher in the 14.0–14.8 m part, lower in the 14.8–15.4 m part, and much lower in the 24.0–26.0 section. All things being equal, an area of elution would have a lower chemical load than an area of infiltration. What we find is that when ion concentrations are high, the elution indices indicate loss, and that when ion concentrations are low (and, indeed, very low), the elution indices indicate no loss. This calls for some other explanation than melt and elution to explain the chemically ‘clean’ part of the ice core.

One other way to examine the chemistry is to investigate the Na^{+}/Cl^{-} ratio. This ratio is taken to be 0.86 in bulk sea water (Bruland 1983). If the precipitation has a Na^{+}/Cl^{-} ratio close to this value, one may assume that the water is proximal to an oceanic source, while lower Na^{+}/Cl^{-} values would indicate an increasing continental influence where continental Cl^{-} may have lowered the ratio. If we analyse the two sections using this index we find that the 14.0–15.4 m section has a Na^{+}/Cl^{-} ratio of 0.67, while the ratio is much lower in the 24.0–26.0 m section. This furthermore suggests that the low chemical content in the 24.0–26.0 m section may represent a more continental environment than what occurred during the deposition of the 14.0–15.4 m section. The high Mg^{2+}/Na^{+} ratio in the 24.0–26.0 m section could accordingly be explained by more continental input of Mg^{2+}. Further the ionic concentrations may have been lower originally in this lower and cleaner ice than in the upper section of the ice core.

We have measured liquid conductivity on nearly the entire core. Although we have measured ionic loading for only *c*. 10% of the ice core, we find that liquid conductivity measurements further argue for a low content of salts in the 23–33.7 m section. The implication is that the low chemical input detected in the 24.0–26.0 m section probably extends to a large part of the ice core (Fig. 4).

To put the chemical composition of this ice core in perspective, we compare with results from Storglaciären and maritime Engabreen (Table 3). We find that ion concentrations in Riukojietna ice are similar in magnitude to ion concentrations in ice from Storglaciären, but are less than an order of magnitude lower than ion concentrations in ice from Engabreen.

In conclusion, we show that there are identifiable chemical signals in Riukojietna ice and that the information may be used to infer environmental changes surrounding the ice cap.

Chronology based on the mass balance reconstruction

It is evident from the ice structural, chemical, and stable isotopic evidence that the Riukojietna ice core contains information of potential use for climatic and environmental interpretations. However, this information can only be meaningfully applied if the ice core temporal framework can be established. Our stratigraphic record lacks obvious chronozones such as ash layers, but is truncated by one or several hiatuses. Because we lack the qualities of a continuous record, we define a crude age model using the mass balance reconstruction and some simple numerical considerations.

We have previously concluded that one hiatus may span the 1910–1988 period based on mass balance considerations (cf. Fig. 3c). Section 1, which contains the firn and the 1995–96 winter snow equalling 1.5 m w.e. in total, only partly accounts for the mass accumulated since 1988 for the height interval 1440–1456 m a.s.l. of 3.3 m w.e. Hence, the *clear ice* facies in section 2 is probably partly composed of ice superimposed since 1988. We estimate that the 3.3 m w.e. boundary (accumulation since 1988) in the ice core is located at approximately 5.3 m depth. The 1988 horizon is therefore likely located somewhere between the base of the firn section, at 3.4 m, and 5.3 m depth.

The period 1910–1988 was a period characterized by overall negative net balances. During this period any accumulation of ice above the **equilibrium line altitude** (**ELA**) likely occurred as superimposed ice. We therefore suggest that section 2 represents post–1910 ice.

Following this, sections 3 and 4 are probably composed of pre-20th century ice (*c*. AD 1600–1910). This conclusion is supported by a reconstruction of the mass balance of Storglaciären by Raper *et al*. (1996). They infer that the last period of sustained positive net mass balance on Storglaciären occurred at the end of the 19th century. This coincides with our reconstructed period with positive net balance at Riukojietna (Fig. 3c). An even earlier period with prevailing positive mass balance on Storglaciären was inferred to have occurred between the first half of the 17th century and the mid–18th century (Raper *et al*. 1996). Tentatively we suggest that sections 3 and 4 represent one or both of the inferred periods of positive net mass balance for Storglaciären during the LIA.

We further suggest that the ice in section 4 is considerably older than the ice in section 3, and that this boundary represents a hiatus. This is because the chemical signals are different and because the amplitude of the δ^{18}O is much more homogenized in section 4. The homogenization of isotopic cycles is primarily a function of time and density (Johnsen *et al*. 2000). Another reason for homogenization may be infiltration of water. However, because the melt index in section 4 is low, the small isotopic amplitude seen in section 4 (24.0–26.0 m, Fig. 4) supports our conclusion that this ice is several centuries old.

Time scale modelling

Estimates of ice core ages can also be based on calculations of the compression of ice. Nye (1963) provided the rate of compression for ice layers close to an ice divide as $l_z = l_s\,(z / H)$, where l is the layer thickness, H is the ice thickness in m w.e. and subscripts s and z represent the values at the surface and at a height z (m w.e.) above the bed. This simple model shows that compression increases linearly with depth assuming isotropic conditions. From Nye's compression rate we can calculate the time, as $t = -(H / a)\ln(z/H)$, where a is the accumulation rate (in m w.e. a^{-1}) (Dansgaard and Johnsen 1969).

Using these simple equations, the age of an ice layer can be estimated if information about ice thickness and the average accumulation rate exists. Past ice thickness and ambient ice depth for a layer are unknown parameters in our case. Although we know the ice thickness in 1996 and ice thickness changes between 1960 and 1978 (Rosqvist and Østrem 1989), we have no information of ice surface changes further back in time. Paterson (1994) suggested that $H = L^{1/2}$, where L is the radius, or length of an ice cap. We can use this relation to estimate how much H has changed between 1910 and 1996. We know from measurements of the position of the glacier snout that it retreated 800 m between 1910 and 1978. We estimate that an additional retreat of 100 m occurred between 1978 and 1996. The distance L between the ice divide and the ice margin at the terminal moraine in 1910 was 2.6 km. Accordingly, L decreased by *c*. 35% until 1996. If we assume $H_{1996} = 80$ m then $H_{1910} = 1.35^{1/2} \times 80 =$ *c*. 93 m, that is, the ice at the drill site was likely *c*. 13 m thicker in 1910.

The average accumulation rate is the second unknown parameter. If we assume that the δ^{18}O variability discovered in some ice core sections indeed reveals annual layers, we can use layer thicknesses to estimate past accumulation rates. For this purpose we use 10 cycles that show higher amplitude than the analytical uncertainty identified in the sections between 11.9–15.4 and between 24.0–26.0 m (19.8–21.3 m; Table 4, Fig. 5). After these cycles/layers become decompressed using Nye's compression equation, they provide the average accumulation rate for that interval (Pohjola *et al*. 2002a). These rates (Table 4) are similar to the present-day accumulation rate (Fig. 3a). Approximations for a and H are then used to calculate the age of the different ice core sections (Table 5). In this operation we use H_{1910} (93 m or 84 m w.e.) because we assume that all ice below 11 m depth accumulated prior to 1910. The results indicate that the boundary between sections 2 and 3 is between 80 and 120 years old and the boundary between sections 3 and 4 between 130 and 190 years old. Although these estimates should be interpreted with caution, they nevertheless correlate surprisingly well with age estimates based on reconstructed mass balance and structural records. As there is additionally *c*. 40 m of ice below the

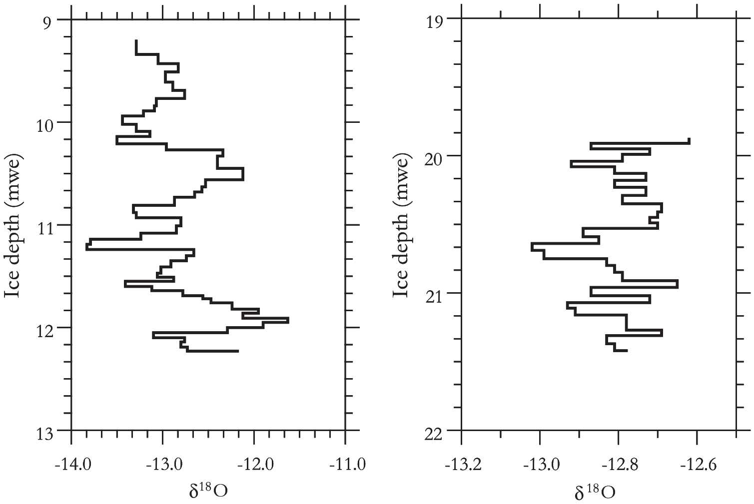

Fig. 5. Higher vertical resolution of the $\delta^{18}O$ distributions in the 11.9–15.4 m and the 24.0–26.0 m intervals discussed with relevance to annual resolution.

recovered ice we tentatively estimate that the ice at the base of the ice cap may be older than one millennium.

Discussion

Based on the ice core stratigraphy we conclude that section 1 (upper 3.4 m), which is characterized by a thin firn layer covering massive clear ice facies, likely represents the period of mass accumulation starting at the end of the1980s (cf. Chinn *et al.* 2005). The dominance of the clear ice facies in section 2 indicates that conditions typical for a wet-snow/superimposed ice zone (Paterson 1994) existed and we propose that 7.1 m w.e. has accumulated as superimposed ice at the drill site between 1910 and 1988. Hence, we infer that, although the ice cap at large lost mass during the 20th Century, accumulation of superimposed ice occurred above the average ELA at the drill site.

In sections 3 and 4, (between 11.0 and 33.7 m) bubbly ice is interspersed with layers of clear ice of varying quantity. During the time of formation conditions were likely colder and snowmelt did not completely fill the pores of the firn, but, instead, formed ice lenses; these processes occur in the percolation zone.

The $\delta^{18}O$ record indicates that the average $\delta^{18}O$ distribution in the ice core during the LIA was similar to the annual average $\delta^{18}O$ in precipitation today. Chemical analyses show that there is a signature in the ice and that ion loading and ion ratios of LIA ice can be used to contrast periods of continental from maritime conditions. Since the chemical analyses were restricted to two short core sections, we used liquid electrical conductivity measurements to extrapolate chemical ion concentrations in the core, assuming that higher conductivity values represent ice containing relatively more salt. Together with the melt index, which shows lower melt below 11 m, the chemical evidence indicates that the ice in sections 3 and 4 (11.0–33.7 m) accumulated during more continental (drier and colder) conditions. The higher chemical concentrations in the 14.0–14.8 m interval, as compared to the 14.8–15.4 m and 24.0–26.0 m intervals, may be explained by an increase in anthropogenically derived ions.

Table 4. Average accumulation rate using the decompressed wavelength from the ice core $\delta^{18}O$ record. The average accumulation rate for each interval is decompressed using the method from Nye (1963).

Depth interval (m w.e.)	Identified cycles	Average compressed accumulation rate (m w.e. yr^{-1})	Average decompressed accumulation rate (m w.e. yr^{-1})
9.38–12.17	10[1]	0.28	0.36
19.83–21.34	10[1]	0.15	0.21
9.38–12.17	9[2]	0.31	0.40
19.83–21.34	7[2]	0.22	0.30
9.38–12.17	5[3]	0.56	0.72

[1] All cycles in the interval.

[2] All cycles where the double amplitude >0.1‰, the analytic uncertainty of the $\delta^{18}O$ measurements.

[3] Using only the larger wavelength cycles in the upper interval (Fig. 5).

Table 5. Age / depth relation calculated on the Riukojietna ice core, using the Nye-timescale. $H = 84$ mwe, a is the accumulation rate and z is the height above the bed.

Ice core (mwe) depth	z (mwe)	t (years) a=0.30 (mwe)	t (years) a=0.25 (mwe)	t (years) a=0.20 (mwe)
8[1]	63	80	97	121
18[2]	53	129	155	193
29[3]	42	194	233	291
65[4]	6	739	887	1108

1. The elevation of the top of section 3
2. The elevation of the top of section 4.
3. The elevation of the bottom of the ice core.
4. Basal part of the ice cap, not ice cored.

The presence of hiatuses caused by periods of negative mass balance poses a problem for age model constructions of Scandinavian ice cores. We employed crude but straightforward techniques to arrive at a temporal framework, and we suggest that such simple techniques may, together with chronozones such as tephra layers, in the future provide some age control. Our results provide useful information concerning the absolute values of and the variability in atmospheric chemistry and oxygen isotopes during the climatologically different period of the LIA. Hence, continued studies of Scandinavian ice cores holds the promise to potentially yield pre-industrial benchmark values of environmental parameters.

Conclusions

Below a top layer of snow and ice that accumulated since 1988 and an intermediate layer of superimposed ice that accumulated between 1910 and 1988 (down to 11.0 m depth), we found ice that accumulated during colder and more continental climatic conditions.

Reconstruction of the mass balance at the ice coring site indicated that this ice accumulated prior to 1910, i.e. during the Little Ice Age (LIA). Nye age-modeling confirms that this ice is >80–120 years old.

Analysis of $\delta^{18}O$ and ion chemistry show that the LIA ice likely contains preserved annual signals. The LIA average $\delta^{18}O$ is similar to the 20th century $\delta^{18}O$. From the chemical record we infer more continental conditions.

Our ice core archive provides benchmark values on atmospheric chemistry and $\delta^{18}O$ for the LIA. This information can be used to study maritime–continental gradients over the region.

We have shown a potential for other Scandinavian ice caps to veil pre-20th century ice with detectable climate signals.

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