A high-altitude snow chemistry record from Amundsenisen, Dronning Maud Land, Antarctica

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ABSTRACT. In this paper a detailed record of major ions from a 20 m deep firn core from Amundsenisen, western Dronning Maud Land, Antarctica, is presented. The core was drilled at $75°$ S, $2°$ E (2900 m a.s.l.) during austral summer 1991/92. The following ions were measured at 3 cm resolution: Na^+, Mg^{2+}, Ca^{2+}, Cl^-, NO_3^-, SO_4^{2-} and CH_3SO_3H (MSA). The core was dated back to 1865 using a combination of chemical records and volcanic reference horizons. The volcanic eruptions identified in this core are Mount Ngauruhoe, New Zealand (1974–75), Mount Agung, Indonesia (1963), Azul, Argentina (1932), and a broad peak that corresponds in time to Tarawera, New Zealand (1886), Falcon Island, South Shetlands, Southern Ocean (1885), and Krakatau, Indonesia (1883). There are no trends in any of the ion records, but the annual to decadal changes are large. The mean concentrations of the measured ions are in agreement with those from other high-altitude cores from the Antarctic plateau. At this core site there may be a correspondence between peaks in the MSA record and major El Niño–Southern Oscillation events.

1. INTRODUCTION

During recent decades a steadily increasing number of studies on the chemical composition of snow and ice on polar ice sheets have been performed in order to understand environmental changes. The variations of the impurity concentrations recorded in the snow are commonly interpreted as changes in the atmospheric environment, despite the fact that the link between the concentrations in the air and those found in the snow and ice is not well understood (e.g. Wolff and Bales, 1996). Many of the ice-core studies have focused largely on a few deep-drilling sites, but we need to improve our understanding of the spatial distribution of chemical species before we can resolve some of the questions involving origin, transport and deposition of the impurities in the Antarctic snow.

The interior of Antarctica is of particular interest because of its unique combination of remoteness from other continents, which excludes to a large degree any local anthropogenic influences, and its extreme atmospheric and meteorological conditions. In recent years, Dronning Maud Land has been the focus of increased attention because it has been selected as a deep-drilling site for the European Project for Ice Coring in Antarctica (EPICA). Several studies from this project concerning accumulation and oxygen isotopes in the area have recently been published (Isaksson and others, 1999; Oerter and others, 1999, 2000; Van den Broeke and others, 1999; Holmlund and others, 2000; Karlöf and others, 2000). Previous studies on the polar plateau include several Swedish Antarctic Research Programme (SWEDARP) expeditions (Isaksson and Karlén, 1994a, b; Isaksson and others, 1996) and one International Trans-Antarctic Scientific

Expedition (ITASE) traverse (Mayewski and Goodwin, 1999) under SWEDARP (Richardson and others, 1997; Stenberg and others, 1998).

Presently only a few glacio-chemical studies have been published from western Dronning Maud Land. The first was performed on the Riiser-Larsen Ice Shelf during the Norwegian Antarctic Research Expedition 1978/79 (Gjessing, 1984), while the more recent studies concern the spatial variation of major ions in snow pits along transects from the coast and inland (Stenberg and others, 1998; M. Stenberg and M. Hansson, unpublished information). These studies report high concentrations of marine-derived ions (Na^+, Mg^{2+}, Ca^{2+}, Cl^-, K^+, MSA) in coastal areas, with a rapid decrease further inland, while the non-sea-salt ions (NO_3^-, $nssSO_4^{2-}$) were not a function of the distance to the coast. Recently, there have also been results from high-resolution continuous flow analysis (CFA) covering the past 2000 years from the polar plateau presented (Sommer and others, 2000b).

In this paper we present chemical data from a 20 m firn core obtained on Amundsenisen in the area around $75°$ S, $2°$ E at 2900 m a.s.l. during SWEDARP 1991/92 (Fig. 1). At this site, Isaksson and others (1996) estimated the mean annual accumulation to be 7.7 cm w.e., and the mean annual temperature, as indicated by the 10 m borehole temperature, to be about $-43.8°$C. This paper focuses on the glacio-chemical record from this core, which is estimated to cover the period 1865–1991. The accumulation and oxygen isotope record are discussed in more detail in Isaksson and others (1996). To summarize, the $\delta^{18}O$ record shows an increasing trend equivalent to a temperature increase of $0.8°$C since 1865, of the same order of magnitude as the calculated overall warming of the Southern Hemisphere (Briffa and Jones,
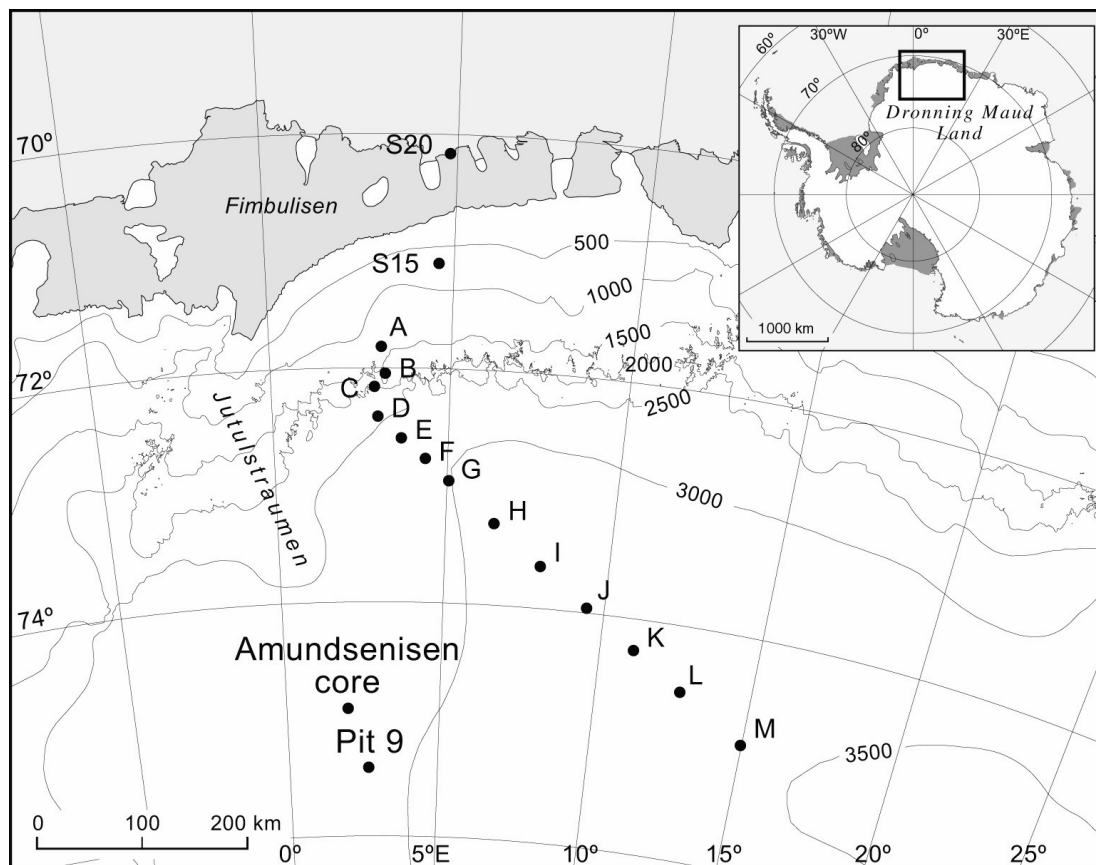
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Fig. 1. Map of the area on Amundsenisen discussed in this paper. Indicated on the map are the core site discussed in this paper, one snow-pit location from ITASE 1993/94 used for comparison and the traverse line with the coring locations from the Norwegian Antarctic Research Expedition (NARE) 1996/97.

1993). The accumulation of 7.7 cm w.e. a^{-1} appears to be rather stable throughout the record and does not reveal any clear trend over the 1865–1991 time period (Isaksson and others, 1996). More recent studies in the same area have also found very stable accumulation rates through time although the annual mean is somewhat higher at this particular site (Oerter and others, 1999, 2000; Karlöf and others, 2000; Sommer and others, 2000a).

2. SAMPLING AND ANALYTICAL PROCEDURES

The core was drilled using a Polar Ice Coring Office (PICO) lightweight coring system with a 3 in (7.6 cm) diameter barrel. The snow/firn density was determined immediately after retrieval by measuring and weighing each core section on an electronic scale. The core was packed in plastic bags, sealed and stored in insulated wooden boxes before further transportation to a freezer on board the ship to Sweden and later to facilities at Stockholm University.

The core was transported frozen from Stockholm to the University of New Hampshire where it was subsampled at 3 cm resolution for major anions, cations and methane-sulphonic acid (MSA). About 2 cm of the outer core section was mechanically cut away to clean the core. It was processed in a cold room (−10°C), using special procedures to minimize contamination (Buck and others, 1992). The sampled core sections were put in pre-cleaned and dried containers. The samples were melted in these containers and aliquots were taken for major-ion, MSA and oxygen isotope analysis. The major ions were analyzed within 2 hours of melting; the MSA and δ^{18}O samples were refrozen. The cations were ana-

lyzed on a CS12 column, and the anions on an AS4A-SC 2 mm column. In both cases, suppressed chromatography was used. MSA was later analyzed on an AS4 column using suppressed conductivity. The methods, including the accuracy of the measurements, are described in Whitlow and others (1992).

3. DATING

The total number of samples per year varied between 3 and 10. An annual stratigraphy was obtained by examining seasonal variations in the sodium and nitrate records (Fig. 2). Comparing these records to the seasonal variation of oxygen isotopes suggests nitrate peaks during summer and sodium during winter, in agreement with other sites on the polar plateau (Legrand and Delmas, 1984). The seasonal cycles in the oxygen isotope record can be observed in the uppermost 4 m, where the densification processes cause diffusion of the signal (Johnsen, 1977). In order to avoid using occasional double peaks as dividing lines for years, we used two ions, which have different seasonal input, to determine the different years. Part of a year's accumulation is likely to be missing due to uneven distribution of snow and wind erosion through the year. The volcanic record interpreted from the non-sea-salt sulphate record provided additional dating reference horizons (Fig. 2) (see section 4.2). We estimate the dating of the core to be ±3 years; however, in the oldest part of the core, before the Krakatau eruption, age control is lacking, so the dating errors there could be about ±5 years. The core was found to cover the period 1865–1991.
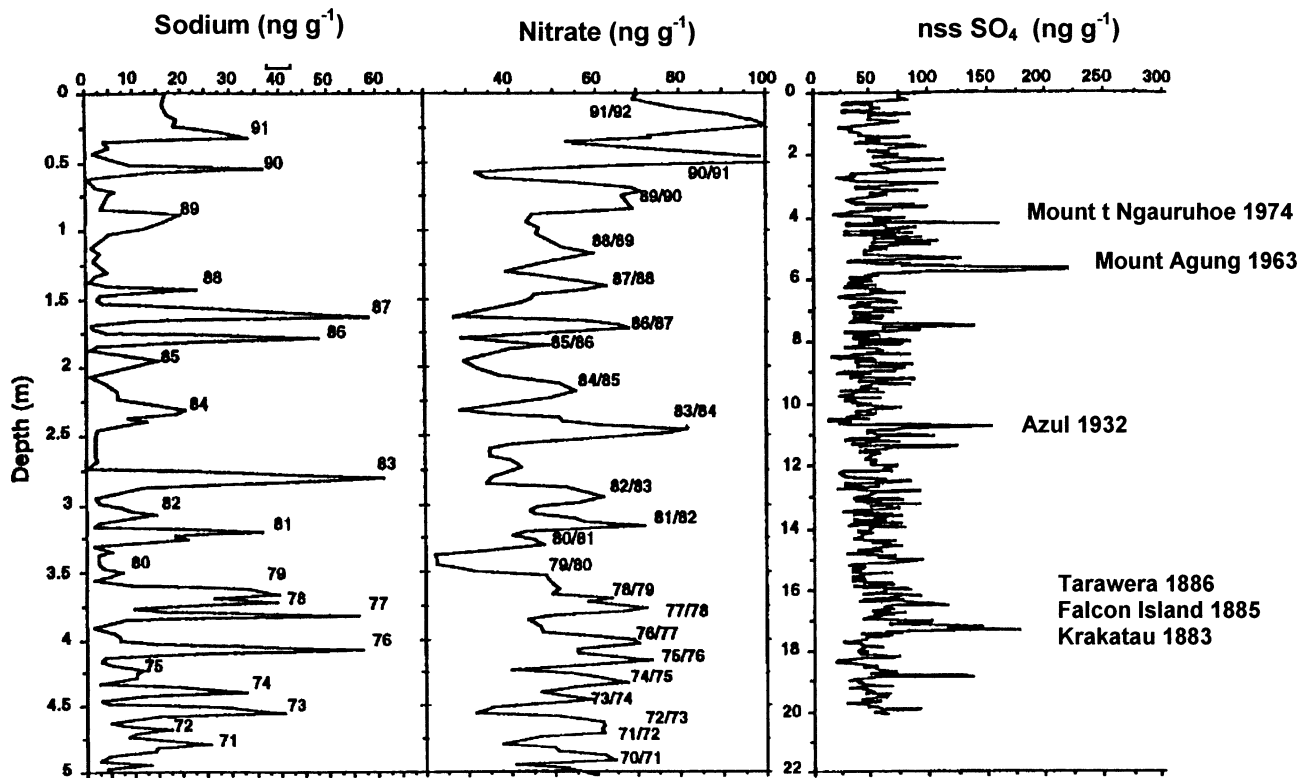
Fig. 2. The upper 5 m, comprising 21 years, of the sodium and nitrate profiles, which were used to obtain an annual time-scale for the core together with the complete profile of the sulphate record from the Dronning Maud Land snow core (from Isaksson and others, 1996). The proposed volcanic peaks are listed in Table 2.

4. ION RECORDS

4.1. Sea salts

In our core, 77% of the magnesium, 59% of the chloride, 43% of the calcium and 6% of the sulphate are of sea-salt origin, in agreement with snow-pit data from the same area (Stenberg and others, 1998). The calculation of the minimum sea-salt contribution follows the method described in Stenberg and others (1998), using Na^+ as the sea-salt indicator following Legrand and Delmas (1988).

Mean concentrations of Na^+, Cl^- and Mg^{2+} are similar to those reported from other high-altitude areas in East Antarctica (Table 1). These major sea-salt components have seasonal variations that are the result of different amounts of input during an annual cycle. If we use the oxygen isotope record as an indicator of the seasons during an annual cycle, then all three ions peak during the time when $\delta^{18}O$ values

Table 1. Concentrations of all the analyzed ions for the Dronning Maud Land core site, estimated to cover the time 1865–1991

Site	Na^+	Mg^{2+}	nss Mg^{2+}	Cl^-	excess Cl^-	SO_4^{2-}	$nssSO_4^{2-}$	Ca^{2+}	$nssCa^{2+}$	No_3^-	MSA	Cl^-/Na^+	Mean annual acc.
	ng g^{-1}	ng g^{-1}	ng g^{-1}	ng g^{-1}	ng g^{-1}	ng g^{-1}	ng g^{-1}	ng g^{-1}	ng g^{-1}	ng g^{-1}	ng g^{-1}	ng g^{-1}	cm w.e.
DML core													
Mean	13.5	2.0	0.3	35.6	11.1	61.6	58.1	1.2	0.7	54.4	7.3	2.8	7.7
Std dev.	14.2	1.8	0.4	24.6	7.8	27.7	27.3	1.3	1.1	14.1	4.9		
Min.	0.3	0.1	−0.8	4.1	−10.4	14.6	12.9	0.2	−0.1	20.2	0.8		
Max.	111.3	13.9	2.9	192.6	55.1	266.7	264.6	20.5	19.6	170.3	44.4		
% of total			23		41		94		57				
Pit 9 ITASE[1]													
1987–93	19.2	2.6	0.8	34.7	11.7	98.4	95			57.7	7.6	2.1	5
South Pole[2]													
1955–88	11.1	2.8		34.1		62.2		4.2		98.8	7–13[3]	2.9[4]	8.5
Dome C[4]													
1760–1980	22			69	48	48				15		3.9	3.4
D 80[4,5]													
1959–69	10			26		37	34			55		2.9	23.7

Note: Data from a nearby snow pit (Stenberg and others, 1998), South Pole (2835 m a.s.l.; Whitlow and others, 1992), Dome C (3240 m a.s.l.; Legrand and Delmas, 1988) and D 80 in Terre Adélie (2525 m a.s.l.; Legrand and Delmas, 1985, 1988).

[1] Stenberg and others (1998). [2] Whitlow and others (1992). [3] Legrand and others (1992). [4] Legrand and others (1988). [5] Legrand and others (1985).

are lowest, i.e. during winter. This has also been shown in a pit study in the same area (Stenberg and others, 1998) and by CFA on several cores from this part of the polar plateau (Sommer and others, 2000b). Aerosol measurements at the South Pole (Cunningham and Zoller, 1981; Bodhaine and others, 1986; Tuncel and others, 1989) clearly show that high input of sea-salt elements is most frequent during the winter storms. The timing of sea-salt input contrasts with that at coastal sites such as Neumayer, 500 km north of the Amundsenisen coring site, where sea salt peaks in the late summer to autumn, at the time of minimum sea-ice extent (Wagenbach and others, 1988).

Temporal variations in sea salts are thought to be associated with changing intensity in surface wind speeds near the ocean surface (e.g. Petit and others, 1981). The temporal variability of the sea-salt species through this record is not large, except for several high concentration spikes through the 1880s and 1890s, (Fig. 3). The most pronounced peaks appear during the winters of 1894 and 1895, and coincide with the Thompson Island eruption which is believed by Lamb (1967) to have occurred in 1895–96 at high southern latitude (54° S, 5° E). This eruption is thought to have destroyed an island and most likely injected large amounts of sea salts into the atmosphere. Legrand and Delmas (1988) suggested that this eruption explained sea-salt peaks around 1894 in a core from Dome C, but there have been no reports from other sites in Antarctica.

The excess Cl$^-$ component is 41%, and a reaction involving sea salts and acid aerosols producing gaseous HCl has been proposed as the major source for excess Cl$^-$ in Antarctica (Legrand and Delmas, 1988). The reaction is more efficient when weather conditions are calm, usually during summer, and the higher the Cl/Na ratio (the indicator for excess Cl$^-$), the stronger the fractionation. Both the Cl/Na ratio of 2.8 in our core (Table 1) and a summer peak agree with the current knowledge of excess Cl$^-$.

Throughout most of our record there are minor fluctuations of the Cl/Na ratio around the mean value. However, during the late 1950s a gradual increase began, culminating in a significant rise around 1980 (Fig. 3), suggesting that the content of HCl in the atmosphere has changed. We have no explanation for this and have not seen it reported elsewhere, although one possibility might be changes in the fractionation of sea salts due to long-range transport.

4.2. Sulphate

Sulphate is generally the dominant ion in high-altitude Antarctic snow, and comes in the form of atmospheric H$_2$SO$_4$, mainly from marine biogenic emission, with sporadic input from volcanic activity. In this core the sea-salt part of sulphate is <6% of the total sulphur (Table 1), and the concentrations show summer maxima, all in agreement with previous studies from the polar plateau (Delmas and Boutron, 1978; Delmas, 1982; Legrand and Delmas, 1984). A well-developed seasonal cycle with summer maxima is often present both in aerosol measurements and in the snow on the polar plateau. The explanation for this could be summer intrusion of stratospheric air masses in combination with enhanced vertical mixing of the troposphere during the summer (Cunningham and Zoller, 1981), or could involve photochemical processes (Legrand and Delmas, 1984).
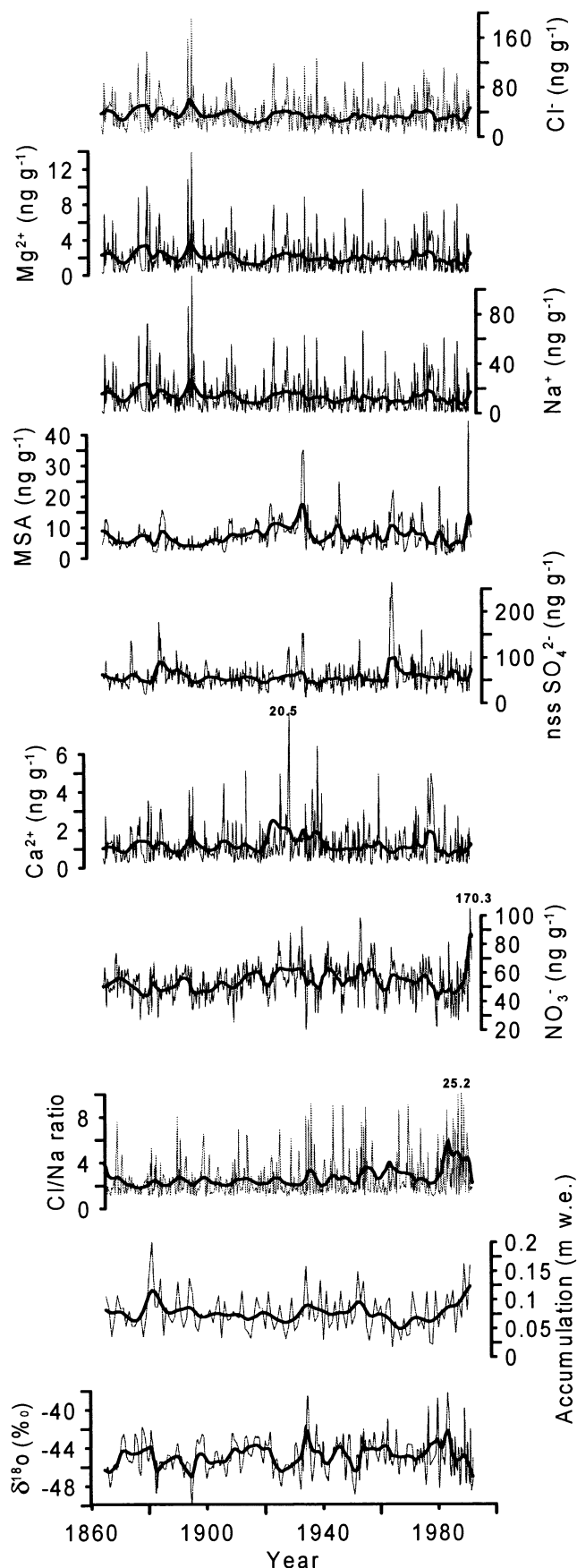

Fig. 3. The ion-concentration records of chloride, magnesium, sodium, MSA, non-sea-salt sulphate, calcium, nitrate and the chloride/sodium ratio, in ng g^{-1}. The oxygen isotope record and accumulation records are included for comparison. For each record the raw data are plotted together with smoothed data using Gaussian weighting coefficients (approximately equivalent to a 7 year moving average).

Table 2. Volcanic eruptions as identified in the non-sea-salt sulphate record

Core depth m	Year	Volcano	Eruption year	Position	Source
4.22	1974–75	Mount Ngauruhoe	1974–75	37° S, 174° E	Aristarain and others (1982) Delmas and others (1985) Mosley-Thompson and others (1991)
5.78	1963–65	Mount Agung	1963	8° S, 115° E	Legrand and Delmas (1984) Delmas and others (1985) Oerter and others (1999)
10.79	1933	Azul	1932	36° S, 71° W	Delmas and others (1985) Karlöf and others (2000)
17.35	1884–86	Krakatua Falcon Island Tarawera	1883 1885 1886	8° S, 106° W	Delmas and others (1985) Karlöf and others (2000)

The most pronounced sulphur peak occurs in summer 1964/65 (Fig. 2; Table 2), corresponding to the eruption of Mount Agung, Indonesia (1963), in good agreement with what is known about the intensity of this eruption (Delmas and others, 1985; Legrand and Delmas, 1987). Other peaks are interpreted as being due to the eruptions of Mount Ngauruhoe, New Zealand (1974), and Azul, Argentina (1932), and one broad peak coincides with the eruptions of Tarawera, New Zealand (1886), Falcon Island, South Shetlands, Southern Ocean (1885), and Krakatau, Indonesia (1883) (Fig. 2; Table 2). The latter two eruptions appear as a clear double peak in areas with higher accumulation (Cole-Dai and others, 1997). Subsequent studies in the same area confirm some of the eruptions (Karlöf and others, 2000; Oerter and others 2000).

4.3. MSA

MSA (CH_3SO_3H) and sulphur dioxide are two products of marine biogenic dimethyl sulphide (DMS; $(CH_3)_2S$) emission. Sulphur dioxide is converted to non-sea-salt SO_4^{2-} in the atmosphere, but MSA does not appear to have any other source than DMS (Lovelock and others, 1972). Therefore, MSA is frequently used as an indicator of the strength of the marine biogenic source (e.g. Ivey and others, 1986; Saigne and Legrand, 1987).

The mean MSA content in this record (Table 1) is comparable to other data collected from the snow on the Antarctic plateau (Legrand and others, 1992; Stenberg and others, 1998). The MSA record does not show any pronounced seasonal variations. The lack of seasonality in MSA has been reported previously from other inland sites in Antarctica (Ivey and others 1986; Legrand and others, 1992), in contrast to coastal areas where seasonal cycles are clearly present in both snow (Mulvaney and others, 1992; Welch and others, 1993; Stenberg and others, 1998) and air samples (Wagenbach, 1996).

The MSA record shows a number of pronounced peaks throughout the record (Fig. 3). Legrand and Feniet-Saigne (1991) also find such peaks in a South Pole core, which they correlate with El Niño–Southern Oscillation (ENSO) events. This has been confirmed by more recent studies from South Pole by Dibb and Whitlow (1996) and Meyerson and others (in press). In our record we have 17 pronounced MSA peaks, which seem to match with ENSO events as listed by Quinn (1992) (Fig. 4; Table 3). Quinn (1992) reports 34 large-scale ENSO events between 1864 and 1987. Some of these occur close together, and are therefore difficult to separate,

and at least two of them comprise more than one event (these peaks are marked 1864, 1866 and 1929–31, 1932 in Fig. 4). The largest peaks are assigned to the ENSO events of 1991–95 and 1929–31, 1932, neither of which was unusually strong, which suggests that the implied link to ENSO is not simple. The MSA record has high concentrations in the uppermost snow layers, corresponding to 1991–92, which could be due to the beginning of the extended ENSO event of 1991–95 (Trenberth and Hoar, 1996). In several cases, the MSA peaks correspond well with the timing of ENSO events, but with a dating error of as much as ±3 years, there is a risk of false correlation since the average frequency of ENSO during this time period is 3.6 years (Quinn, 1992). When more records become available through EPICA, a stacked MSA record produced by several snow cores may provide different results.

Several shallow-pit studies in Dronning Maud Land have also clearly indicated increased MSA concentrations during recent ENSO events (Stenberg and others, 1998; M.

Table 3. The ENSO chronology together with the corresponding MSA peak in the 20 m snow core

ENSO event year	MSA peak year	Rating
1991–95*	1991	
1986–87	1986–87	M
1982–83	1981–83	VS
1976–77	1974–75	M
1972–73	1971–72	S+
1968–69	1968	M–
1965–66	1963–65	S
1957–58	1957–59	S
1943–44	1944–46	M
1932	1932–35	M+
1929–31		M+
1925–26	1925–26	S
1923	1922–23	M
1918–20	1917–18	S+
1907	1908–09	M+
1884–85	1884–86	M+
1876–78	1877–79	VS
1865–66	1865–66	M+
1864		S+

Notes: Both the chronology and the rate of strength for each ENSO event are according to Quinn (1992). The rating system is VS, very strong; S, strong; M, medium. The proposed corresponding ENSO event is indicated for each MSA peak in Figure 4.

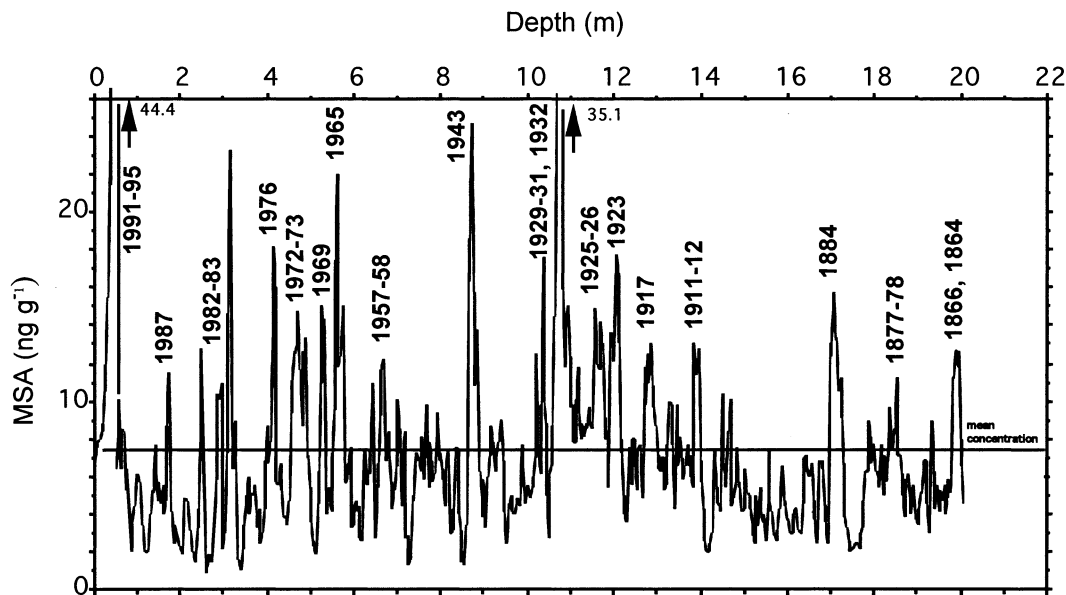* Defined after Trenberth and Hoar (1996).

Fig. 4. The concentration record of MSA from the Amundsenisen ice core. The MSA peaks are correlated with ENSO events, which are indicated by years, after Quinn (1992). The corresponding ENSO events are listed in Table 3 together with other relevant information.

Stenberg and M. Hansson, unpublished information). These results suggest that the pronounced MSA peaks are both spatially reproduced and preserved down the snowpack.

Furthermore, although the amplitudes of the corresponding ENSO events in our record from Amundsenisen and the South Pole core (Legrand and Feniet-Saigne, 1991) do not correspond very well, both records show low concentrations and a lack of major peaks between the mid-1940s and late 1950s, a period with little ENSO activity according to Quinn (1992).

The high MSA concentrations during some of the ENSO years suggest that this is more than a regional phenomenon and that the large-scale circulation pattern is affected. White and Petersen (1996) presented intriguing evidence of changes associated with the Antarctic continent on interannual time-scales. Using data concerning atmospheric pressure at sea level, wind stress, sea surface temperature and sea-ice extent over the Southern Ocean, they track a system of coupled anomalies. During a period of 4–5 years this system of anomalies, which they call the Antarctic Circumpolar Wave, moves eastwards around the continent. They suggest that its initiation is associated with El Niño in the equatorial Pacific. If the atmosphere and ocean are coupled in such a stable way, it may be reasonable to infer the MSA–ENSO link presented here.

The implications for the source area of MSA are discussed by Legrand and others (1992) by using the difference in timing between coastal and high-altitude sites to infer a low- or mid-latitude source for the high-altitude areas, and a local source for the coastal zones of Antarctica. They base their arguments on the difference in the MSA/nssSO$_4^{2-}$ molar ratio (R), which is latitude-dependent, high-latitude areas having a higher value (Berresheim, 1987; Bates and others, 1992). The climatic implications of changing MSA content in snow/ice cores are focused particularly on the value of R, which is supposed to have a direct correlation to the air temperature (Legrand and others, 1992). The average value of the MSA/nssSO$_4^{2-}$ molar ratio for this core is 0.125, corresponding to a source at low or mid-latitudes.

There are significant increases in R during ENSO events, also found by Legrand and Feniet-Saigne (1991) in the South Pole record. They suggested an enhanced contribution from high-latitude sources during ENSO events. Some of the ENSO events have a corresponding nssSO$_4^{2-}$ peak, but not

all. This implies either that the non-sea-salt sulphate does not arrive simultaneously with MSA or that it has additional, as yet unknown, sources. It thus appears that MSA consists of a mixture of local and distant material, the former attracted by the relative proximity to the coast (about 500 km) and the latter by the high altitude.

Stenberg and others (1998) found that MSA and sulphate have different spatial distributions in this area, implying that their sources are different and that the molar ratio R should be used with great caution. Other studies of the spatial variability in MSA suggest that Antarctic high-altitude sites have an additional source of MSA (Kreutz and Mayewski, 1999), perhaps through the upper troposphere (Legrand and Feniet-Saigne, 1991).

4.4. Non-sea-salt calcium: a crustal indicator

The non-sea-salt calcium is considered to be primarily of continental origin and is therefore used as an indicator of changes in wind speed over continental areas, and of changes in transportation strength (Mayewski and Lyons, 1982; De Angelis and others, 1987).

The non-sea-salt calcium fraction is 57% of the total calcium content in this core (Table 1), which is somewhat higher than reported from the same area over the past 2000 years (Sommer and others, 2000b). The record shows seasonal variations, with maxima during the austral summer, in agreement with results from CFA by Sommer and others (2000b). Aerosol measurements at the South Pole (Tuncel and others, 1989) show high concentrations of crustal material in the summer, which gradually decrease to a minimum during the austral winter. Hogan (1984) observed that terrestrial aerosols at South Pole are associated with the upper-troposphere or the lower-stratosphere source layer. The seasonal distribution is largely determined by the strength of the cyclonic wind barrier around the continent and the temperature inversion layer, both of which are stronger during winter. Thus changes in input of dust at high altitudes in Antarctica have been interpreted as changes in the summer upper-tropospheric wind.

In general, the calcium record indicates an overall increase between 1865 and 1940 and a decrease thereafter, but the record also consists of cycles, which seem to have a

periodicity of 10–20 years. Particularly strong peaks or increases occur during the periods 1880–85, 1905–12, 1920–40, 1960 and 1975–77, with calcium increases of >15 times the mean value. This is in general agreement with the peaks in the sea-salt record. These peaks correspond to periods with lower $\delta^{18}O$ content, i.e. colder periods (Fig. 3). The most pronounced calcium increase takes place between 1920 and 1940 and is associated with a cold period.

4.5. Nitrate

The nitrate record from Amundsenisen has a total concentration of 54.5 ng g^{-1} (Table 1), lower than the South Pole concentration but much higher than for most other sites on the polar plateau (Legrand and Delmas, 1986, 1988). The nitrate record shows clear seasonal variations (Fig. 3), with summer peaks, as at other high-altitude sites (e.g. Wolff, 1995). The highest concentrations are found in the uppermost 0.6 m, most likely because of gaseous interaction with the atmosphere through the porous snow (density around 0.35 g cm^{-3}), as at many low-accumulation sites in Antarctica (e.g. Mayewski and Legrand, 1990; Dibb and Whitlow, 1996; M. Stenberg and M. Hansson, unpublished information). The source for nitrate in Antarctica is widely debated (see Wolff (1995) for a review of this topic).

5. CONCLUSIONS

None of the ion records from this high-altitude firn core from western Dronning Maud Land show any trend over the time period 1865–1991. The MSA concentrations show a substantial increase in connection with ENSO years. This correspondence is not perfect throughout the core; whether this is due to dating problems or mechanisms within the teleconnection is difficult to say. Results from White and Petersen (1996) strongly suggest an ENSO influence in both the oceans and atmosphere around Antarctica. In addition, a recent meteorological study focused on western Dronning Maud Land using model precipitation data suggests that the variability in accumulation is directly linked to ENSO, so one can expect this signal to be seen in ice cores from this area (Noone and others, 1999).

Shallow-snow-pit studies of the spatial variations of major ions, including MSA (Stenberg and others, 1998; Stenberg and Hansson, in press), suggest that this core site is representative also for a larger part of this area of the polar plateau. Due to the extensive EPICA pre-site investigations, many more shallow and medium–deep ice cores have recently been drilled in the same area, and the results so far available suggest that the spatial variability is large and redistribution by wind may cause problems with correlation of short time periods (Sommer and others, 2000a, b). At this point there are no other published MSA records from this area that cover the time period discussed in this paper. However, more chemical data will soon become available from this area on the polar plateau, which will provide a more complete picture of glacio-chemical conditions.

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REFERENCES

Aristarain, A. J., R. J. Delmas and M. Briat. 1982. Snow chemistry on James Ross Island (Antarctic Peninsula). *J. Geophys. Res.*, **87**(C13), 11,004–11,012.

Bates, T. S., J. A. Calhoun and P. K. Quinn. 1992. Variations in the methane-sulfonate to sulfate molar ratio in submicrometer marine aerosol particles over the south Pacific Ocean. *J. Geophys. Res.*, **97**(D9), 9859–9865.

Berresheim, H. 1987. Biogenic sulfur emissions from the subantarctic and Antarctic oceans. *J. Geophys. Res.*, **92**(D11), 13,245–13,262.

Bodhaine, B. A., J. J. Deluisi, J. M. Harris, P. Houmere and S. Bauman. 1986. Aerosol measurements at the South Pole. *Tellus*, **38B**(3–4), 223–235.

Briffa, K. R. and P. D. Jones. 1993. Global surface air temperature variations during the twentieth century. Part 2: Implications for large-scale high-frequency palaeoclimatic studies. *Holocene*, **3**(1), 77–88.

Buck, C. F., P. A. Mayewski, M. J. Spencer, S. Whitlow, M. S. Twickler and D. Barrett. 1992. Determination of major ions in snow and ice cores by ion chromatography. *J. Chromatogr.*, **594**(1–2), 225–228.

Cole-Dai, J., E. Mosley-Thompson and L. G. Thompson. 1997. Annually resolved Southern Hemisphere volcanic history from two Antarctic ice cores. *J. Geophys. Res.*, **102**(D14), 16,761–16,771.

Cunningham, W. L. and W. H. Zoller. 1981. The chemical composition of remote area aerosols. *J. Aerosol Sci.*, **12**(4), 367–384.

De Angelis, M., N. I. Barkov and V. N. Petrov. 1987. Aerosol concentrations over the last climatic cycle (160 kyr) from an Antarctic ice core. *Nature*, **325**(6102), 318–325.

Delmas, R. J. 1982. Antarctic sulphate budget. *Nature*, **299**, 677–678.

Delmas, R. and C. Boutron. 1978. Sulfate in Antarctic snow: spatio-temporal distribution. *Atmos. Environ.*, **12**, 723–728.

Delmas, R. J., M. Legrand, A. J. Aristarain and F. Zanolini. 1985. Volcanic deposits in Antarctic snow and ice. *J. Geophys. Res.*, **90**(D7), 12,901–12,920.

Dibb, J. E. and S. Whitlow. 1996. Recent climatic anomalies and their impact on snow chemistry at South Pole, 1987–1994. *Geophys. Res. Lett.*, **23**(10), 1115–1118.

Gjessing, Y. 1984. Marine and non-marine composition of the chemical composition of the snow at the Riiser–Larsen Ice Shelf in Antarctica. *Atmos. Environ.*, **18**(4), 825–830.

Hogan, A. W. 1984. Aerosol exchange in the remote troposphere. *Tellus*, **38B**(3–4), 197–213.

Holmlund, P. *and 15 others*. 2000. Spatial gradients in snow layering and 10 m temperatures at two EPICA–Dronning Maud Land (Antarctica) pre-site-survey drill sites. *Ann. Glaciol.*, **30**, 13–19.

Isaksson, E. and W. Karlén. 1994a. High resolution climatic information from short firn cores, western Dronning Maud Land, Antarctica. *Climatic Change*, **26**(4), 421–434.

Isaksson, E. and W. Karlén. 1994b. Spatial and temporal patterns in snow accumulation, western Dronning Maud Land, Antarctica. *J. Glaciol.*, **40**(135), 399–409.

Isaksson, E., W. Karlén, N. Gundestrup, P. Mayewski, S. Whitlow and M. Twickler. 1996. A century of accumulation and temperature changes in Dronning Maud Land, Antarctica. *J. Geophys. Res.*, **101**(D3), 7085–7094.

Isaksson, E., M. R. van den Broeke, J.-G. Winther, L. Karlöf, J. F. Pinglot and N. Gundestrup. 1999. Accumulation and proxy-temperature variability in Dronning Maud Land, Antarctica, determined from shallow firn cores. *Ann. Glaciol.*, **29**, 17–22.

Ivey, J. P., D. M. Davies, V. Morgan and G. P. Ayers. 1986. Methanesulphonate in Antarctic ice. *Tellus*, **38B**(5), 375–379.

Johnsen, S. J. 1977. Stable isotope homogenization of polar firn and ice. *International Association of Hydrological Sciences Publication* 118 (Symposium at Grenoble 1975 — *Isotopes and Impurities in Snow and Ice*), 210–219.

Karlöf, L. *and 12 others*. 2000. A 1500 year record of accumulation at Amundsenisen, western Dronning Maud Land, Antarctica, derived from electrical and radioactive measurements on a 120 m ice core. *J. Geophys. Res.*, **105**(D10), 12,471–12,483.

Kreutz, K. J. and P. A. Mayewski. 1999. Spatial variability of Antarctic surface snow glaciochemistry: implications for paleoatmospheric circulation reconstructions. *Antarct. Sci.*, **11**(1), 105–118.

Lamb, H. H. 1967. The problem of Thompson Island: volcanic eruptions and meteorological evidence. *Br. Antarct. Surv. Bull.* 13, 85–88.

Legrand, M. R. and R. J. Delmas. 1984. The ionic balance of Antarctic snow: a 10 year detailed record. *Atmos. Environ.*, **18**(9), 1867–1874.

Legrand, M. and R. J. Delmas. 1985. Spatial and temporal variations of snow chemistry in Terre Adélie (East Antarctica). *Ann. Glaciol.*, **7**, 20–25.

Legrand, M. and R. J. Delmas. 1986. Relative contributions of tropospheric

and stratospheric sources to nitrate in Antarctic snow. *Tellus*, **38B**(3–4), 236–249.

Legrand, M. R. and R. J. Delmas. 1987. A 220-year continuous record of volcanic H_2SO_4 in the Antarctic ice sheet. *Nature*, **327**(6124), 671–676.

Legrand, M. R. and R. J. Delmas. 1988. Formation of HCl in the Antarctic atmosphere. *J. Geophys. Res.*, **93**(D6), 7153–7168.

Legrand, M. and C. Feniet-Saigne. 1991. Methanesulfonic acid in south polar snow layers: a record of strong El Niño? *Geophys. Res. Lett.*, **18**(2), 187–190.

Legrand, M., C. Feniet-Saigne, E. S. Saltzman and C. Germain. 1992. Spatial and temporal variations of methanesulfonic acid and non sea salt sulfate in Antarctic ice. *J. Atmos. Chem.*, **14**(1–4), 245–260.

Lovelock, J. E., R. J. Maggs and R. A. Rasmussen. 1972. Atmospheric dimethyl sulphide and the natural sulphur cycle. *Nature*, **237**, 452–453.

Mayewski, P. A. and I. Goodwin. 1999. Antarctic's role pursued in global climate change. *Eos*, **80**(35), 398–400.

Mayewski, P. A. and M. Legrand. 1990. Recent increase in nitrate concentration of Antarctic snow. *Nature*, **346**(6281), 258–260.

Mayewski, P. A. and W. B. Lyons. 1982. Source and climatic implication of the reactive iron and reactive silicate concentrations found in a core from Meserve Glacier, Antarctica. *Geophys. Res. Lett.*, **9**(3), 190–192.

Meyerson, E. A., P. A. Mayewski, K. J. Kreutz, L. D. Meeker, S. I. Whitlow and M. S. Twickler. In press. The polar expression of ENSO and sea-ice variability as recorded in a South Pole ice core. *Ann. Glaciol.*, **35**.

Mosley-Thompson, E., J. Dai, L. G. Thompson, P. M. Grootes, J. K. Arbogast and J. F. Paskievitch. 1991. Glaciological studies at Siple Station (Antarctica): potential ice-core paleoclimatic record. *J. Glaciol.*, **37**(125), 11–22.

Mulvaney, R., E. C. Pasteur, D. A. Peel, E. S. Saltzman and P.-Y. Whung. 1992. The ratio of MSA to non-sea-salt sulphate in Antarctic Peninsula ice cores. *Tellus*, **44B**(4), 295–303.

Noone, D., J. Turner and R. Mulvaney. 1999. Atmospheric signals and characteristics of accumulation in Dronning Maud Land, Antarctica. *J. Geophys. Res.*, **104**(D16), 19,191–19,211.

Oerter, H., W. Graf, F. Wilhelms, A. Minikin and H. Miller. 1999. Accumulation studies on Amundsenisen, Dronning Maud Land, by means of tritium, dielectric profiling and stable-isotope measurements: first results from the 1995–96 and 1996–97 field seasons. *Ann. Glaciol.*, **29**, 1–9.

Oerter, H. and 6 others. 2000. Accumulation rates in Dronning Maud Land, Antarctica, as revealed by dielectric-profiling measurements of shallow firn cores. *Ann. Glaciol.*, **30**, 27–34.

Petit, J.-R., M. Briat and A. Royer. 1981. Ice age aerosol content from East Antarctic ice core samples and past wind strength. *Nature*, **293**(5831), 391–394.

Quinn, W. H. 1992. A study of Southern Oscillation-related climatic activity for A.D. 622–1990 incorporating Nile River flood data. *In* Diaz, H. F. and V. Markgraf, *eds. El Niño. Historical and paleoclimatic aspects of the Southern Oscillation.* Cambridge, Cambridge University Press, 119–149.

Richardson, C., E. Aarholt, S.-E. Hamran, P. Holmlund and E. Isaksson. 1997. Spatial distribution of snow in western Dronning Maud Land, East Antarctica, mapped by a ground-based snow radar. *J. Geophys. Res.*, **102**(B9), 20,343–20,353.

Saigne, C. and M. Legrand. 1987. Measurements of methanesulphonic acid in Antarctic ice. *Nature*, **330**(6145), 240–242.

Sommer, S. and 9 others. 2000a. Glacio-chemical study spanning the past 2 kyr on three ice cores from Dronning Maud Land, Antarctica. 1. Annually resolved accumulation rates. *J. Geophys. Res.*, **105**(D24), 29,411–29,421.

Sommer, S., D. Wagenbach, R. Mulvaney and H. Fischer. 2000b. Glacio-chemical study spanning the past 2 kyr on three ice cores from Dronning Maud Land, Antarctica. 2. Seasonally resolved chemical records. *J. Geophys. Res.*, **105**(D24), 29,423–29,433.

Stenberg, M. and 7 others. 1998. Spatial variability of snow chemistry in western Dronning Maud Land, Antarctica. *Ann. Glaciol.*, **27**, 378–384.

Trenberth, K. E. and T. J. Hoar. 1996. The 1990–1995 El Niño–Southern Oscillation event: longest on record. *Geophys. Res. Lett.*, **23**(1), 57–60.

Tuncel, G., N. K. Aras and W. H. Zoller. 1989. Temporal variations and sources of elements in the South Pole atmosphere. 1. Nonenriched and moderately enriched elements. *J. Geophys. Res.*, **94**(D10), 13,025–13,038.

Van den Broeke, M. R. and 6 others. 1999. Climate variables along a traverse line in Dronning Maud Land, East Antarctica. *J. Glaciol.*, **45**(150), 295–302.

Wagenbach, D. 1996. Coastal Antarctica: atmospheric chemical composition and atmospheric transport. *In* Wolff, E. W. and R. C. Bales, *eds. Chemical exchange between the atmosphere and polar snow.* Berlin, etc., Springer-Verlag, 173–199. (NATO ASI Series I: Global Environmental Change 43.)

Wagenbach, D., U. Görlach, K. Moser and K. O. Münnich. 1988. Coastal Antarctic aerosol: the seasonal pattern of its chemical composition. *Tellus*, **40B**(5), 426–436.

Welch, K. A., P. A. Mayewski and S. I. Whitlow. 1993. Methanesulfonic acid in coastal Antarctic snow related to sea ice extent. *Geophys. Res. Lett.*, **20**(6), 443–446.

White, W. B. and R. G. Peterson. 1996. An Antarctic circumpolar wave in surface pressure, wind, temperature and sea-ice extent. *Nature*, **380**(6576), 699–702.

Whitlow, S., P. A. Mayewski and J. E. Dibb. 1992. A comparison of major chemical species seasonal concentration and accumulation at the South Pole and Summit, Greenland. *Atmos. Environ.*, **26A**(11), 2045–2054.

Wolff, E. W. 1995. Nitrate in polar ice. *In* Delmas, R. J., *ed. Ice core studies of global biogeochemical cycles.* Berlin, etc., Springer-Verlag, 195–224. (NATO ASI Series I: Global Environmental Change 30.)

Wolff, E. W. and R. C. Bales, *eds.* 1996. *Chemical exchange between the atmosphere and polar snow.* Berlin, etc., Springer-Verlag. NATO Advanced Science Institutes. (NATO ASI Series I: Global Environmental Change 43.)