Accumulation and proxy-temperature variability in Dronning Maud Land, Antarctica, determined from shallow firn cores

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ABSTRACT. During 1996–97 a European Project for Ice Coring in Antarctica (EPICA) pre-site surveying traverse worked in the area between 70° S, 5° E and 75° S, 15° E in Dronning Maud Land. We present data obtained from 10 and 20 m deep firn cores drilled between the coast and 600 km inland (to 3450 m a.s.l.). The cores were analyzed for electrical conductivity measurements and total β activity to obtain accumulation data between known time horizons. In addition, some of the cores were analyzed for oxygen isotopes. Annual accumulation varies from 271 mm w.e. at Fimbulisen to 24 mm w.e. at 2840 m a.s.l. Accumulation at core sites 2400–3000 m a.s.l. has increased by 16–48% since 1965 compared to the 1955–65 period. However, the core sites above 3250 m a.s.l. and one core location on the ice shelf show a decrease during the same period. Furthermore, no change can be detected at the most inland site for the period 1815–1996. In all the cores the last few years seem to have been some of the warmest in these records.

INTRODUCTION

Natural climate variability over much of Antarctica is still poorly known, both because Antarctic instrumental records are short (starting in 1956–57) and because such records are restricted to the coastal zones. Longer records are available from the Antarctic Peninsula and South Orkney Islands, but these show little correspondence with records from continental Antarctica and therefore cannot be used to extend the short instrumental records (Raper and others, 1984).

Ice-core records remain the most valuable tool for deriving past climate data in Antarctica. By selecting a suitable drill location it is possible to retrieve information extending back several hundred thousand years (e.g. Jouzel and others, 1993). At the other extreme, it is possible to collect high-resolution information on seasonal time-scales with cores from high accumulation areas (e.g. Morgan and Van Ommen, 1997). The well-established relation between air temperature and snowfall isotope chemistry permits us to use ice cores to extend the instrumental records back in time and expand the spatial coverage.

Nevertheless, basic glaciological and climatological data are still lacking for large areas of Antarctica; one such area is the interior of Dronning Maud Land (DML). During the 1996–97 field season the European Project for Ice Coring in Antarctica (EPICA) pre-site surveying traverse, supported by the Norwegian Antarctic Research Expedition (NARE), worked in the area between 70° S, 5° E and 75° S, 15° E in

DML (Fig. 1). The aim of the traverse was to collect glaciological and meteorological data which would help in selecting a drill site for the EPICA DML deep ice core. Details of the traverse and field methods are described in Winther and others (1997). The spatial variability of some of the climate variables collected along the traverse are presented and discussed in Van den Broeke and others (1999).

In this paper we present accumulation and stable-isotope data obtained from some of the 10 and 20 m deep firn cores obtained in 1996–97 between the coast and 600 km inland (to 3450 m a.s.l.) (Fig. 1; Table 1). In the discussion of the data we also include a firn-core record from Amundsenisen (75° S, 2° E at 2900 m a.s.l.; Fig. 1), obtained during the Swedish Antarctic Research Programme in 1991–92 (Isaksson and others, 1996). The Amundsenisen core site is located about 300 km west of core M and should be under similar climatic influences.

DATING

The cores were dated using the known time horizons 1964–65 and 1954–55 from the nuclear bomb tests (Fig. 2). All the cores except S15 and S20 were analyzed for β radioactivity in Grenoble, France, using methods described in Pourchet and others (1983).

Electrical conductivity measurement (ECM) analysis was performed on all the cores. Horizons of increased
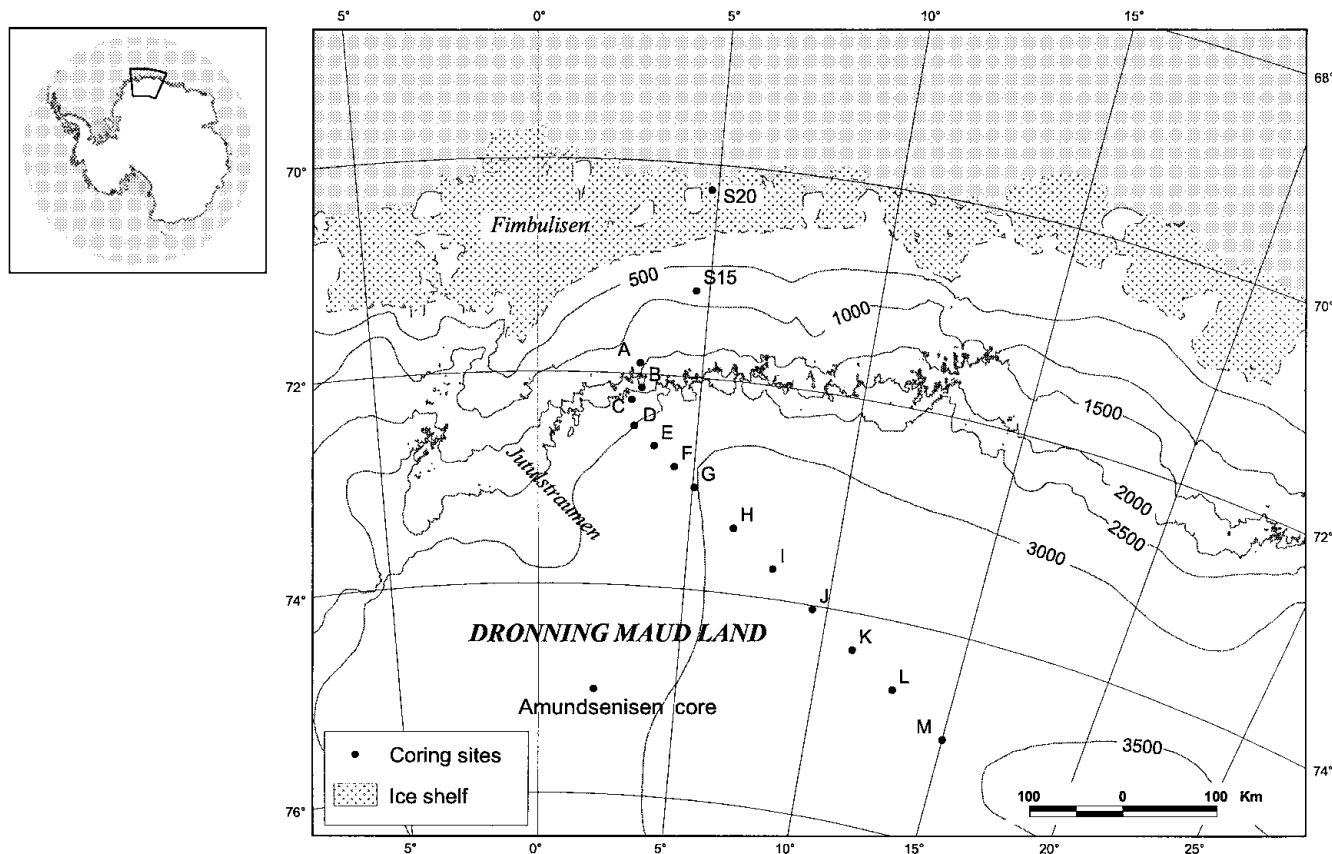
Fig. 1. Map of the field area in eastern DML with the core sites marked. Core-site data are presented in Table 1.
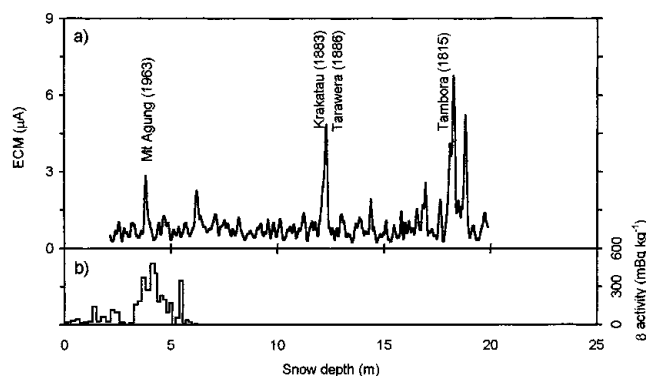

Fig. 2. The ECM (a) and total β activity (b) records from core M. The identified distinctive volcanic eruptions are indicated in the ECM record (running mean).

conductivity were assumed to represent the eruptions of Mount Agung (1963), Krakatau (1883), Tarawera (1886) and Tambora (1815). These reference horizons were used as a complement to the total β activity measurements and for dating the cores beyond the 1954–55 layer (Fig. 2). The ECM records were obtained in a cold room using the method for ice developed by Hammer (1980). Measurements were made under constant-temperature conditions (−15°C) and with a constant applied voltage of 500 V which is lower than in most systems. In our ECM system we found that using higher voltage produced a noisier record and we therefore chose to use the lower setting. A fresh surface was cut in the firn with a microtome knife, and the sliding speed of the electrode was controlled manually. Each core piece was measured several times to ensure the reproducibility of the experiment.

Seven of the cores were analyzed for oxygen isotopes at the University of Copenhagen. Core S20 was sampled with 40–50 mm resolution, S15 with 50 mm and cores A, B, E, L and M with 30 mm. Seasonal variations in the oxygen-isotope stratigraphy and the ECM record are used to date the two coastal cores S15 and S20. Core S20 from the most coastal site at Fimbul Ice Shelf is dated back to 1955.

No reference horizons before the 1954–55 layer could be found in the ECM records from cores E and L. Therefore, dating of cores E and L back to about 1920 was done using the accumulation rate obtained between the radioactive layers and assuming constant accumulation throughout the record. The dating error is estimated to be ±3 years for cores S15 and S20 (based on ambiguities in the $\delta^{18}O$ stratigraphy), ±2 years at the reference horizons for the cores dated with total β activity (based on the sample size and annual accumulation rate) and ±2 years for the volcanic layers in core M dated with ECM. The dating error for the period before 1954–55 in cores E and L may be even larger because of the lack of time control.

All available accumulation and oxygen-isotope data are presented in Table 1 together with the core-site information. In the following discussion of the data, we distinguish between cores from the ice shelf (core S20), the coastal zone (cores S15, A, and B) and the polar plateau (cores E, L and M).

OXYGEN-ISOTOPE DATA

Reliable transfer functions are necessary when translating ice-core records into climate data. The temperature vs $\delta^{18}O$ relationship is empirical (Dansgaard, 1964) and many factors can influence the content of $\delta^{18}O$ in the snow, such as change of moisture source, distance to moisture source and

Table 1. Sites for snow cores including accumulation and oxygen-isotope data

Site	Location		Elevation	T_{10m}	Accumulation 1955–96	Accumulation 1965–96	Accumulation 1955–65	Dating method	Mean $\delta^{18}O$
			m a.s.l.	°C	mm w.e. a^{-1}	mm w.e. a^{-1}	mm w.e. a^{-1}		‰
S$_{20}$	70°14′30″ S	04°48′40″ E[1]	63[2]	−17.5	271 ± 20[6]	265 ± 20	288 ± 20[6]	$\delta^{18}O$	−23.04[7]
S$_{15}$	71°11′30″ S	04°35′50″ E[1]	800[2]	−20.3	–	244 ± 18	–	$\delta^{18}O$	−26.81[7]
A	71°54′00″ S	03°05′00″ E[1]	1520[2]	−25.6[6]	–	135 ± 10	–	β	−33.36[7]
B	72°08′01″ S	03°10′31″ E	2044	−28.1	–	171 ± 13	–	β	−33.40[7]
C	72°15′30″ S	02°53′28″ E	2400	−30.3	119 ± 9	123 ± 9	106 ± 8	β	−33.42[8]
D	72°30′30″ S	03°00′00″ E	2610	−34.8	112 ± 9	116 ± 9	100 ± 8	β	−38.40[8]
E	72°40′30″ S	03°39′46″ E	2751	−36.9[4]	55 ± 5	59 ± 5	40 ± 3	β	−40.42[7]
F	72°51′30″ S	04°21′05″ E	2840	−38.5	23 ± 2	24 ± 2	22 ± 2	β	–
G	73°02′30″ S	05°02′39″ E	2929	−40.1[4]	28 ± 2	30 ± 2	23 ± 2	β	−41.20[8]
H	73°23′30″ S	06°27′38″ E	3074	−42.7[5]	44 ± 4	46 ± 4	38 ± 3	β	–
I	73°43′30″ S	07°56′26″ E	3174	−44.6	52 ± 4	53 ± 4	52 ± 4	β	−44.87[8]
J	74°02′30″ S	09°29′30″ E	3268	−46.3	55 ± 4	52 ± 4	67 ± 5	β	–
K	74°21′30″ S	11°06′13″ E	3341	−47.8[5]	45 ± 3	44 ± 3	49 ± 4	β	−46.32[8]
L	74°38′30″ S	12°47′27″ E	3406	−49.3[5]	45 ± 3	41 ± 3	58 ± 5	β	−47.81[7]
M	74°59′30″ S	15°00′06″ E	3453	−51.3	51 ± 4	45 ± 4	70 ± 6	β	−49.49[7]

[1] Horizontal position from hand-held GPS receivers.

[2] Elevation from tracked vehicle GPS record.

[3] From AWS thermistor string.

[4] Corrected for non-stationarily: −1.1 K.

[5] Measured at 9.8, 9.9 and 9.0 m depth at sites H, K and L, respectively.

[6] Starting year is 1956.

[7] Average over the whole core length.

[8] Average from a 2 m snow pit.

transportation history. Using oxygen-isotope records as proxy-temperature records requires knowledge of the specific temperature–$\delta^{18}O$ relationship for a particular area. Because of the empirical nature of this relationship the temperature–$\delta^{18}O$ gradient can provide insight into some of these processes (see, e.g., Graf and others, 1994). We have established a gradient for this area using 10 m borehole temperatures and the mean $\delta^{18}O$ from shallow pits and cores (altogether 12 sites) along the traverse (Fig. 3). The temperature was measured with a thermistor cable which was left in the borehole overnight (Winther and others, 1997). The gradient for this area is 0.73‰ °C^{-1} ($R^2 = 0.98$) (Fig. 3) which is very close to the gradient of 0.75‰ °C^{-1} proposed for East Antarctica by Lorius and Merlivat (1977). However, since the temperature–$\delta^{18}O$ relationship using this method is a

spatial gradient based on the current conditions, it may not be valid for other time periods, a problem that has been discussed by Jouzel and others (1997). Because of the lack of a better method we choose to use the spatial gradient but with caution.

Polar plateau

We analyzed three cores from the polar plateau (E, L and M) for oxygen isotopes (Fig. 4). However, the low accumulation rate makes it impossible to count annual layers using $\delta^{18}O$ peaks; there are simply too many missing layers. Therefore we have interpolated the time-scale between the stratigraphic marker horizons. The longest record, from core M, extends back beyond the Tambora eruption in 1815 (Fig. 2).

The $\delta^{18}O$ record in core M suggests that temperatures were above average for most of the period 1820–60. A cold period begins before the Krakatau eruption (1883) and continues until the beginning of 1900 (Fig. 4). This cold period is also evident in the core from Amundsenisen, but here it begins at the time of the eruption. Overall, the multi-year patterns in the $\delta^{18}O$ records from these two cores are similar.

There is a pronounced isotopic warm event in cores L and M around 7.5 m, which would correspond approximately to 1930–40 (Table 2), a period known to have been particularly warm from Southern Hemisphere temperature records (Jones and Briffa, 1992). This apparent warming is also seen in the Amundsenisen record, but is less distinct.

Cores E, L and M all indicate that the last few years have been some of the warmest during the past century. There is no temperature record available from the nearby Sanae station after 1992, but our coastal records (which are discussed below) and oxygen-isotope data from western Fimbul Ice Shelf (Melvold, 1999) also indicate a temperature increase during the last 3 years.
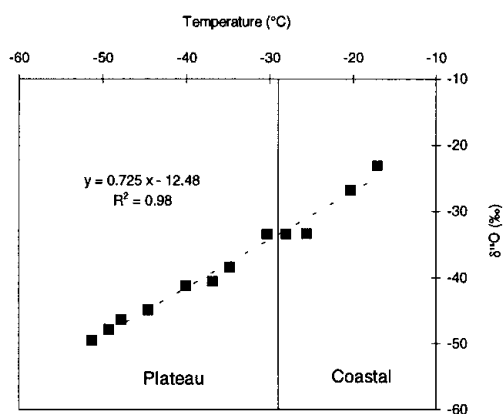

Fig. 3. The mean $\delta^{18}O$ data from 12 of the sites (S20, S15, A, B, C, D, E, G, I, K, L and M) vs the 10 m core-hole temperature along the travel route. Some of the values are the mean from 2 m deep snow pits, and others from 10–20 m cores (see Table 1).
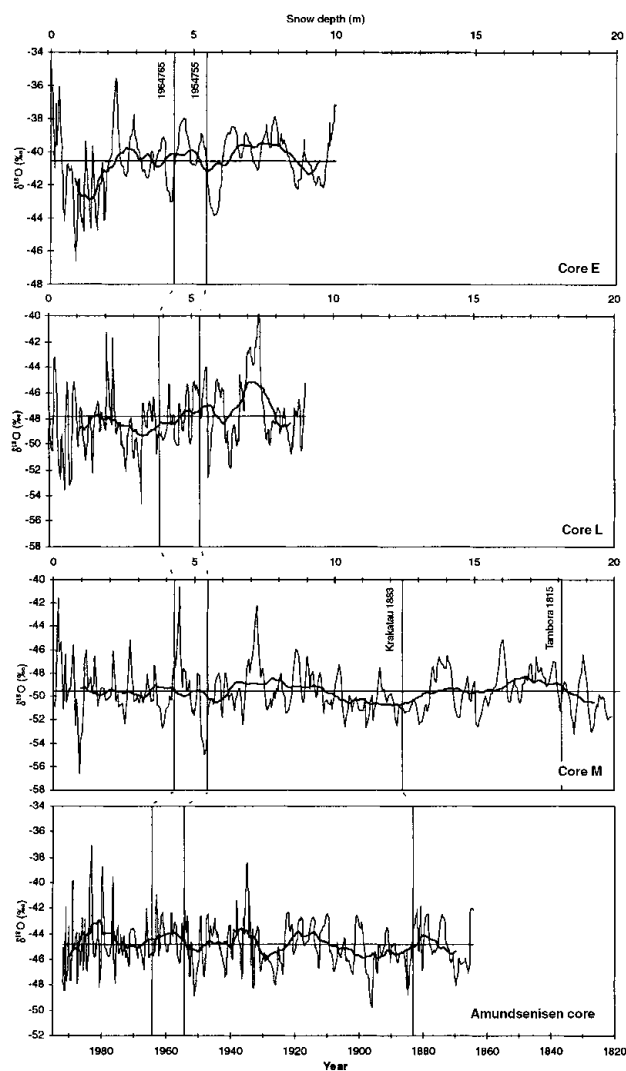
Fig. 4. The $\delta^{18}O$ stratigraphy from four cores from the polar plateau. Cores E, L and M are from this work, and the other $\delta^{18}O$ record is from Amundsenisen, western DML (Isaksson and others, 1996). On the Amundsenisen core, major ions were analyzed and annual layers identified. The time-stratigraphic markers from β activity, ECM and the mean line are indicated. The records have been smoothed with a running mean, equivalent to 3 years.

The oxygen-isotope record from Amundsenisen indicates a temperature increase of about 0.8°C since 1865 (Isaksson and others, 1996). Also the oxygen-isotope record from core M indicates a temperature increase equivalent to

Table 2. $\delta^{18}O$ values for the different time periods established by the ECM and β activity dating methods

Date	E	L	M	Amundsenisen
Core mean	−40.4	−47.8	−49.5	−44.8
1965–96[1]	−41.0	−48.5	−49.5	−44.6
1955–65	−39.9	−47.9	−47.9	−44.2
1930–40[2]	−39.9	−44.9	−47.9	−44.1
1883–1955			−49.7	−45.0
1815–83			−49.3	

[1] 1965–92 for Amundsenisen core.
[2] The time period 1930–40 is based on assuming constant accumulation rate.

almost 1°C since about 1883. This is more than the overall warming trend of 0.5°C in the Southern Hemisphere instrumental records since 1880 (Jones and Briffa, 1992) but in accordance with estimates based on information from early expeditions to and around the continent (Jones, 1990). In a review of Little Ice Age conditions on the Antarctic continent, Mosley-Thompson (1992) did not find a persistent pattern for the whole continent, but both oxygen-isotope and dust data point towards cooler temperatures on the polar plateau during much of the time period referred to as the Little Ice Age (approximately AD 1500–1880).

Coastal zone

Simple statistical analysis does not reveal any significant trends in the oxygen-isotope records from the coastal sites S20, S15, A and B (Fig. 5). Since the coastal climate is influenced by cyclonic intrusions one should not expect a close correlation between the $\delta^{18}O$ records and the temperature record from Sanae. However, cores A, B and S20 all suggest colder temperatures in the early 1960s, in agreement with Sanae temperatures.

ACCUMULATION DATA

Annual accumulation varies from 271 mm w.e. at Fimbul Ice Shelf to 45 mm w.e. at the traverse turning-point at 75° S, 15° E (Table 1; Fig. 1). Accumulation was also measured continuously with snow radar (Richardson and Holmlund, 1999).

In general, the spatial distribution of accumulation, with a rapid decrease of accumulation with altitude and distance from the shelf edge (Fig. 6), is in agreement with data from stake measurements and stratigraphic studies in snow pits collected during the period 1956–60 by Norwegian scientists who overwintered at Norway station on Fimbul Ice Shelf (Lunde, 1961). They determined the annual accumulation at the ice shelf to be 495 mm w.e. for the period 1940–59. Melvold (1999) has estimated 480 mm w.e. for the 1965–93 period.

Polar plateau

The accumulation on the high-altitude plateau varies from 123 mm w.e. a^{-1} in the area just south of the nunatak range to as little as 24 mm w.e. a^{-1} at site F. Van den Broeke and others (1999) use a katabatic-wind model to show that the accumulation minimum is probably due to high wind erosion in this area. Also, ^{137}C deposition rates confirm the effect of wind activity. The accumulation data from the sites above about 3000 m a.s.l. are in agreement with data collected during the third leg of the South Pole–Queen Maud Land Traverse in summer 1967–68, about 200 km southwest of our traverse line (Picciotto and others, 1971).

Calculating accumulation from the total β activity peaks yields significant accumulation decreases on the highest parts of the polar plateau between the time periods 1955–65 and 1965–96, hereafter referred to as P1 and P2, respectively (Fig. 6). This decrease varies between 10% and 35% at core sites J, K, L and M; however, the changes at site K are within the error bars. There is a 15–50% increase between P1 and P2 at core locations C, D, E, G and H, while core sites F and I do not show any changes (Fig. 6). Isotope values in core M suggest that temperatures were about 2°C (1.6‰) lower for P2 than for P1 (Table 2). The same value for core L is <1°C (0.6‰). At site E, where accumulation has in-
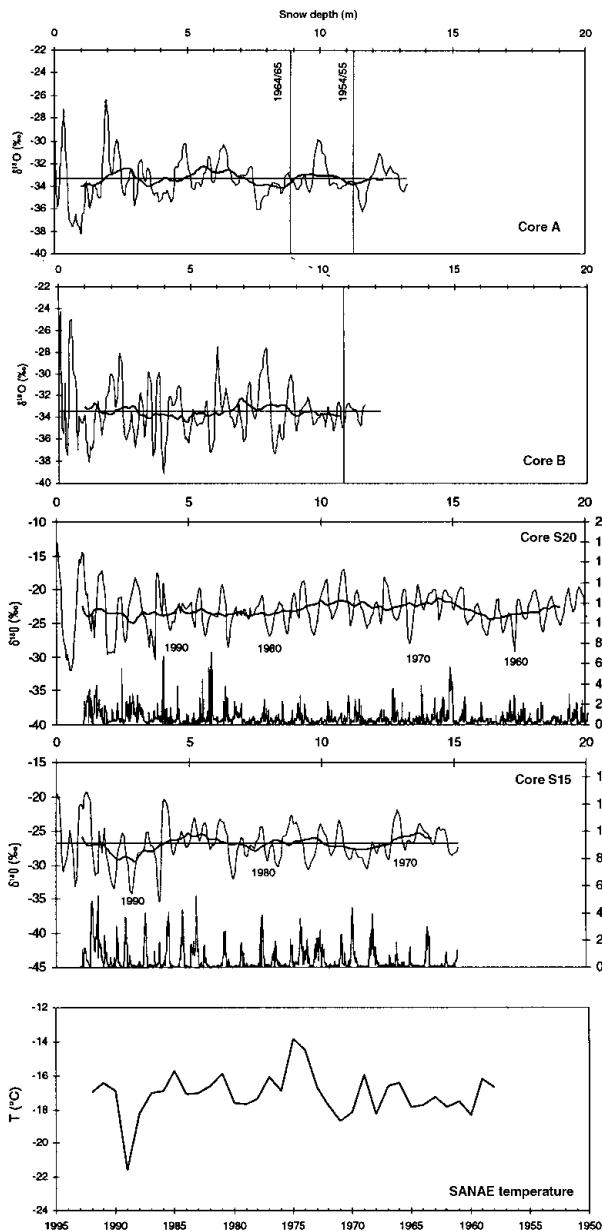
Fig. 5. $\delta^{18}O$ records from four cores from the coastal zone. They have been smoothed with a running mean, equivalent to 3 years. The ECM records for cores S20 and S15 indicate annual variability and provided help with the dating. The time horizons from the total β activity analysis are indicated in the $\delta^{18}O$ records from cores A and B. The mean annual temperature record from Sanae is included for comparison.
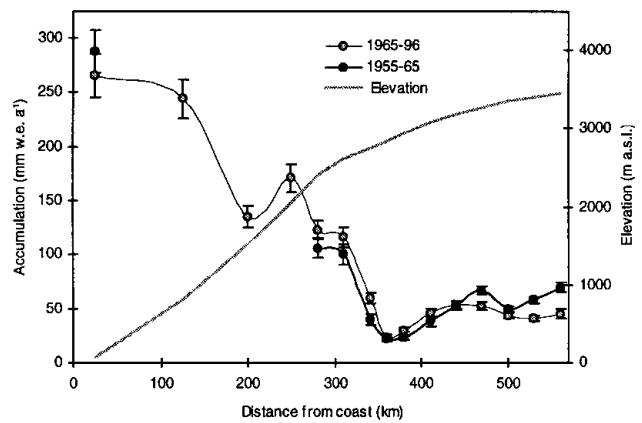

Fig. 6. Spatial and temporal variability of accumulation data along the traverse. The grey squares are the annual accumulation for the period 1965–96, and the black dots for the period 1955–65. The error bars indicate the ±1 standard error. The data suggest an accumulation increase on the polar plateau at about 2400–3000 m a.s.l. and a decrease on the sites above 3250 m. The ice-shelf record shows an accumulation decrease. The grey line is the elevation profile.
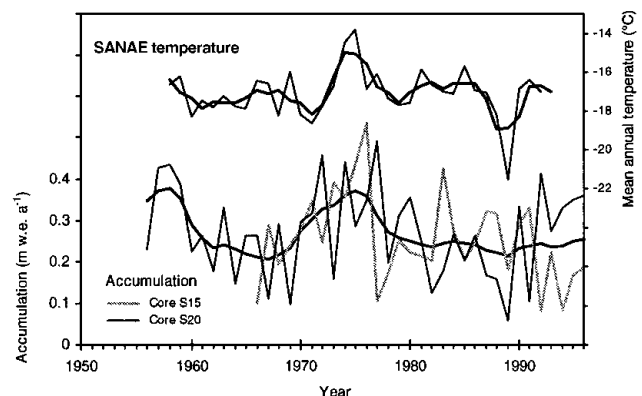

Fig. 7. The annual and the averaged and smoothed (3 year running mean) accumulation record from core S20 on Fimbul Ice Shelf and core S15 from the area just above the grounding line plotted with the smoothed (3 year running mean) temperature record from Sanae (data obtained from South African Weather Bureau).

creased, there is a change of 1.5‰ in mean $\delta^{18}O$ value between P1 and P2.

The time horizons from the volcanic horizons in core M (Fig. 2) provide additional time periods when accumulation can be calculated and compared. Accumulation appears to have been stable over this time period (1815–1996). Despite a large annual fluctuation in accumulation, this relatively well-dated core from Amundsenisen does not show any significant changes since 1865.

Several studies have reported recent accumulation increases from both coastal areas and the polar plateau in East Antarctica (e.g. Petit and others, 1982; Morgan and others, 1991; Mosley-Thompson and others, 1995). However, this study and studies from western DML about 500 km west (Isaksson and Karlén, 1994; Isaksson and others 1996)

suggest that accumulation increases do not necessarily apply to the whole polar plateau.

Coastal zone

Data from the two coastal cores S20 and S15 provide a 40 year long time series of accumulation, based on annual dating from the $\delta^{18}O$ stratigraphy. The averaged and smoothed record of these data shows a distinct positive anomaly for the years 1970–76 (Fig. 7). In addition, the years 1955–59 show higher accumulation than the overall mean in cores S20 and S15. The year-to-year correlation with the Sanae temperature record is not significant, but the multi-year trends are similar. During the particularly cold year 1989, accumulation at both core sites S15 and S20 was <20% lower than the overall 1955–96 mean. During the period 1960–70 the accumulation is below mean in both cores S20 and S15. At Riiser–Larsen Ice Shelf about 1000 km west, the accumulation is low from 1955 to 1975, increases from 1975 to 1985 and decreases up to 1988

(Isaksson and Karlén, 1994). The accumulation record from Riiser–Larsen Ice Shelf corresponds quite closely with the temperature record from Halley.

Data from the Fimbul Ice Shelf around 200 km west and Jutulstraumen also indicate higher accumulation between 1955 and 1965 compared to the period 1965 to present (Melvold, 1999).

CONCLUDING REMARKS

Our finding of a complicated spatial pattern of accumulation is not easy to explain. Furthermore, it demonstrates that the accumulation increases reported from several areas of the polar plateau in East Antarctica are not affecting high-altitude areas over the whole continent. Finding the mechanisms for this is a challenge for the meteorological and modelling communities. This also has important implications for sea-level change in connection with the global climate warming expected in the next century.

The question is, how reliable are accumulation changes as indicators of temperature increase? The fact is that temperature changes cannot explain these large changes in accumulation. Therefore, circulation variations appear to be important (King and Turner, 1997). In comparing data from different areas we must remember that different precipitation mechanisms operate in coastal and inland areas. While coastal areas are mostly affected by cyclones which rarely penetrate far inland, the high-altitude areas have a large input of clear-sky precipitation which could be more directly sensitive to atmospheric temperature changes than to the frequency of cyclone paths. The processes controlling the spatial variability of accumulation are still not understood, and therefore it is essential to improve the observations.

ACKNOWLEDGEMENTS

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