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Ice core melt features in relation to Antarctic coastal climate

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Abstract: Measurement of light intensity transmission was carried out on an ice core S100 from coastal Dronning Maud Land (DML). Ice lenses were observed in digital pictures of the core and recorded as peaks in the light transmittance record. The frequency of ice layer occurrence was compared with climate proxy data (e.g. oxygen isotopes), annual accumulation rate derived from the same ice core, and available meteorological data from coastal stations in DML. The mean annual frequency of melting events remains constant for the last ~150 years. However, fewer melting features are visible at depths corresponding to approximately 1890–1930 AD and the number of ice lenses increases again after 1930 AD. Most years during this period have negative summer temperature anomalies and positive annual accumulation anomalies. The increase in melting frequency around ~1930 AD corresponds to the beginning of a decreasing trend in accumulation and an increasing trend in oxygen isotope record. On annual time scales, a relatively good match exists between ice layer frequencies and mean summer temperatures recorded at nearby meteorological stations (Novolazarevskaya, Sanae, Syowa and Halley) only for some years. There is a poor agreement between melt feature frequencies and oxygen isotope records on longer time scales. Melt layer frequency proved difficult to explain with standard climate data and ice core derived proxies. These results suggest a local character for the melt events and a strong influence of surface topography.

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Key words: climate proxies, ice layers, light transmittance, melting, stratigraphy

Introduction

Antarctic firn and ice cores serve as a very valuable source of information about climate changes and they help to extend the instrumental record both temporally and spatially (e.g. Jouzel *et al*. 1993, Noone *et al*. 1999). However, a thorough knowledge of glaciological processes, as well as of the atmospheric conditions under which a certain snow layer was formed, is essential to obtain an accurate interpretation of the data. Analysing visible layers is one of the oldest stratigraphical techniques (e.g. Benson 1962, Langway 1967). Optical features in snow and firn appear to include an annual (seasonal) signal, as well as longer period signals (Alley *et al*. 1997, Hawley *et al*. 2003) but the origin of these features is not fully understood. If some components relate to changes in the depositional environment beyond seasonality, we might be able to link them to important climatic cycles. Regionally, melting plays an important role in the surface mass balance of Antarctica. Extensive melting has been reported from low-lying ice shelves all around the continent, mostly using remote sensing techniques. Melt ponding has probably led to the disintegration of the northerly ice shelves in the Antarctic Peninsula (Scambos *et al.* 2003). Van den Broeke *et al.* (2004a) calculated melt rates using data from Automatic Weather Stations (AWS) in western Dronning Maud Land [DML] (at 74°S), and found typical values of 20 mm w.e. (water equivalent) of melt per year for the period 1998–2001, distributed over 4–6 events per year. Several other studies also claim an important role of sub-surface melting (Liston & Winther 2005).

Firn core S100 is ~100 m long and was drilled in the summer season 2000/2001 during the joint European Project for Ice Coring in Antarctica (EPICA) and the Norwegian Antarctic Research Expedition (NARE). The drilling site is located in the coastal area of DML, in the eastern part of Fimbulisen at 70°14'S, 04°48'E, and 48 m a.s.l (Fig. 1). The whole S100 core contains climatic information over the last ~260 yrs. It was dated back to ~1737 AD ± 3 yrs by identification of volcanic horizons (Di-Electric Profiling and Electrical Conductivity Measurement) in combination with seasonal layer counting from high resolution oxygen isotope records (Kaczmarska *et al.* 2004). The average accumulation rate over the period ~1737–2000 AD is 0.29 m a^{-1} w.e. A significant (95%) negative trend in the 20th century found in the accumulation record is in agreement with other coastal ice cores from

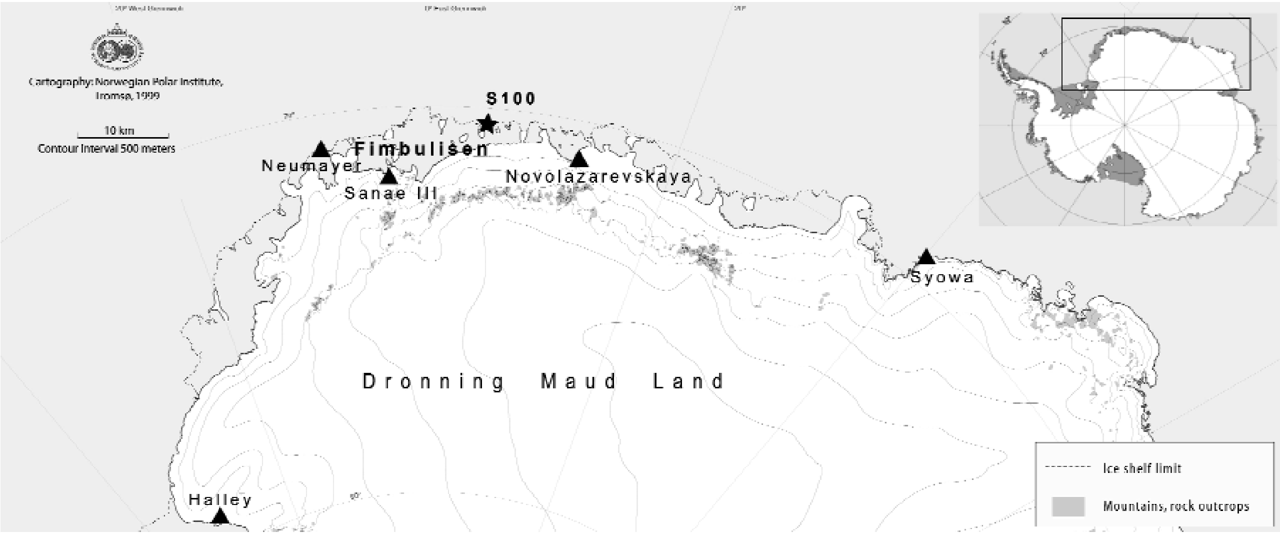

Fig. 1. Location of the drilling site S100 and meteorological stations in coastal Dronning Maud Land, Antarctica: Halley, Neumayer, Sanae, Novolazarevskaya, Syowa.

DML (Melvold 1999, Isaksson *et al.* 1996, Isaksson & Melvold 2002, Schlosser & Oerter 2002). The meteorological data used in this study are from the Halley, Neumayer, Sanae, Novolazarevskaya and Syowa stations. All sites referred to in the paper are marked in Fig. 1.

In this study we discuss the intensity of light transmitted through a coastal Antarctic firn core, S100 (Fig. 1) and recorded on digital images. The goal of this paper is to track the ice layers within the ice core, and investigate to what degree their occurrence is a result of local climate conditions.

Methods

The optical stratigraphy method is based on a measurement of intensity of light transmitted through snow/firn/ice. The S100 optical stratigraphy profile covers the whole core length (depth 0.37–99.8 m). The measurement was carried out in the cold laboratory in the Norwegian Polar Institute, with a room temperature between -15°C and -18°C. The core was cut horizontally, and then the surface was microtomed in order to obtain a clean and flat slab (*c.* 4 cm thick in this core). This slab was put on a light table and the core was digitally photographed (Nikon Coolpix 990 camera) in overlapping intervals. In the light table the light bulb is placed under a transparent fluorescent plate on which the core is placed. After taking the pictures, digital data were transferred to a PC for further processing. The photographs were merged together to form a single dataset covering the whole core. We read out values from ~200 pixel rows across the core (every 0.8 mm depth) and used the average as a measure of transmitted light intensity, yielding a detailed digital optical stratigraphy record from the firn core (Fig. 2). This record was then compared with other data.

The intensity of the transmitted light is influenced by snow/firn/ice density as well as the presence of air bubbles and impurities (Svensson *et al.* 2005). In snow and firn, the

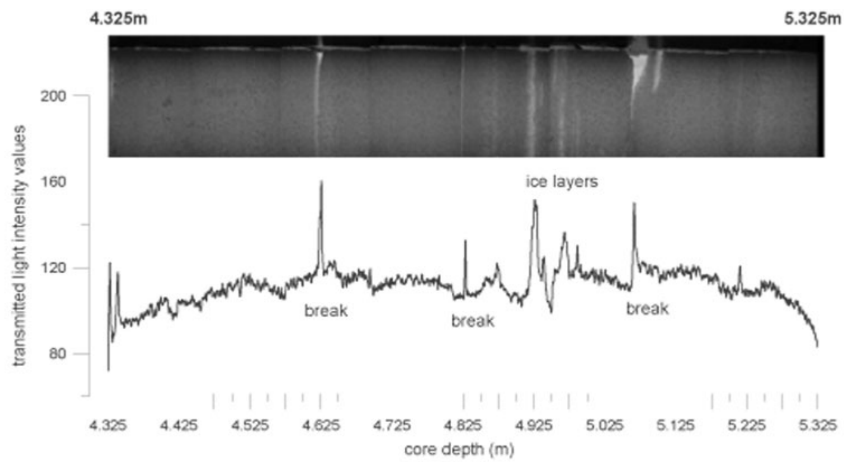

Fig. 2. Example of firn core stratigraphy as recorded by digital camera (upper panel) and the light intensity signal (lower panel) for S100 core at 4.325–5.325 m depth.

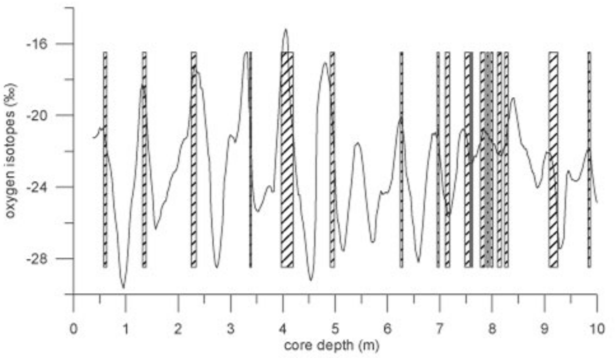

Fig. 3. Location of ice layers with reference to summer (less negative values of $\delta^{18}O$) and winter peaks (more negative values of $\delta^{18}O$) in the oxygen isotope record (depth 0–10 m) derived from the S100 firn core.

intensity of transmitted light is relatively low because of strong light scattering and subsequent absorption in the material. Therefore, these areas appear dark in the images. Peak transmittance occurs in places with ice lens breaks in the core where scattering and absorption are weaker. That is why ice layers are obvious optical features that can easily be identified by eye in digital pictures. They also have corresponding peaks in the light intensity record (Fig. 2). However, since breaks in the core create similar peaks in the transmitted light record, the light intensity signal should be analysed together with digital images.

An ice layer is defined as a nearly bubble-free core section (Alley & Anandakrishnan 1995) occupying a large portion (*c.* 75% or more) of the cross-section of the core, with snow crystals clustered and usually bigger than in other parts of the core. In this paper, the terms 'melt layers', 'ice layers', and 'ice lenses' are used to denote the same. We interpreted ice lenses as a result of melting and refreezing during the warmest part of a year (Alley & Anandakrishnan 1995) thus showing a seasonal pattern (Fig. 3). One difficulty is that summer meltwater may percolate into firn from the winter or from a previous year (Benson 1962). As a result, a summer ice layer may be located below real summer strata as indicated by the $\delta^{18}O$ record.

Choosing a visual approach to identify ice layers in the core helped to avoid counting 'false' peaks in the record (e.g. core break peaks) as ice lenses. Two people counted ice layers (recorded on digital images) independently several times and the result of each counting remained

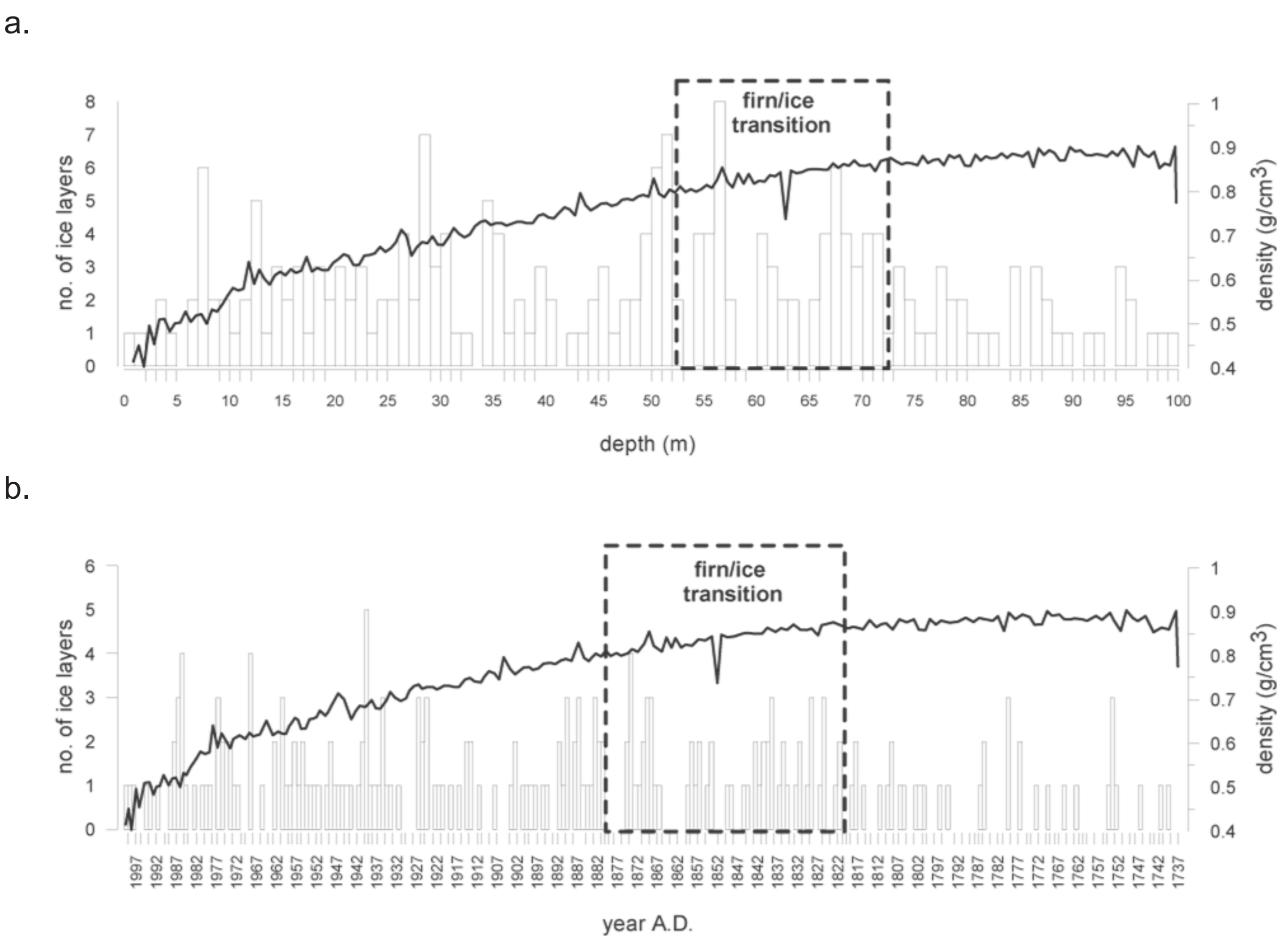

Fig. 4. Frequency of ice layers **a.** per 1 m depth, and **b.** per year as identified in the core S100. We also show the firn density curve.

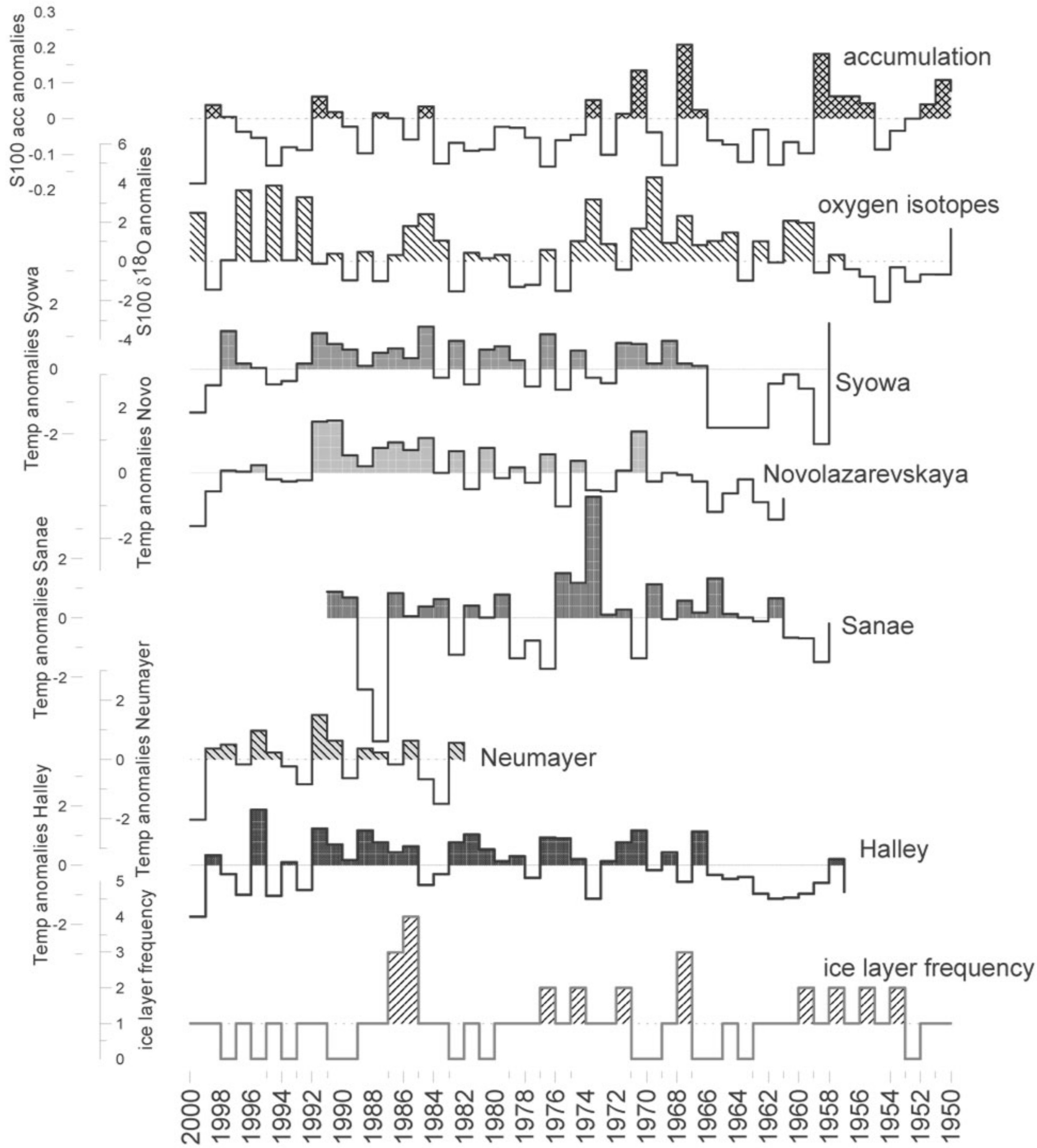

Fig. 5. Comparison between ice core-derived climate proxy data and meteorological records. Graphs from top to bottom: S100 annual accumulation anomalies, S100 annual $\delta^{18}O$ anomalies and mean summer (December–February) air temperature anomalies from coastal meteorological stations: Sanae (meteorological data obtained from the South African Weather Bureau), Syowa, Novolazarevskaya, Neumayer and Halley (data for Syowa, Novolazarevskaya, Neumayer and Halley available from BAS web page: http://www.antarctica.ac.uk/met/gjma/temps.html) and S100 annual melt layers frequency.

within 219 ± 5 layers.

We present stratigraphy data for the whole core; however, we focused on the upper section of the core (0–25 m depth) for the comparison with recent meteorological data. The firn/ice transition zone, starting with a density of ~0.83 g cm^{-3} (Paterson 1994), is reached in this core between 53 and 73 m depth. This depth interval corresponds to the period 1820–80 AD.

We used the depth-age scale based on dating by other methods (Kaczmarska *et al.* 2004) to calculate the frequency of melt layers per year (Fig. 4a & b). The summer is plotted as 2000 AD in all figures.

Results and discussion

We found 219 ± 5 melt layers along the S100 core (Fig. 4).

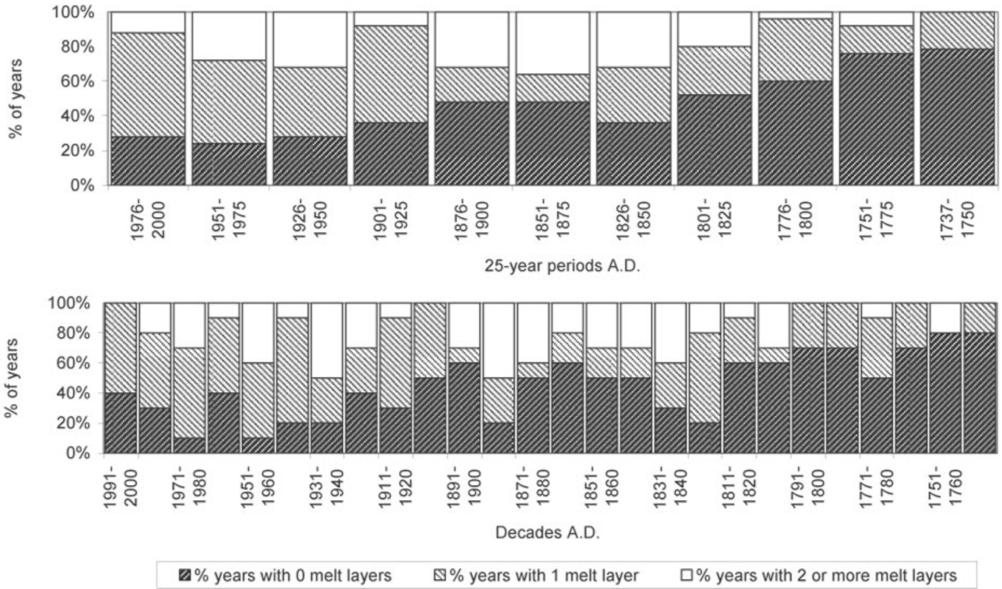

Fig. 6. Percentage of years with 0 ice layers (white), 1 ice layer (grey) and 2 or more ice layers (black) as derived from the S100 ice core for **a.** decades, and **b.** 25-year periods.

Some sections of the core show several ice layers clustered together while in other places only single and relatively thin yet distinctive layers are found (Fig. 3).

We used data from the five nearest meteorological stations: Halley, Neumayer, Sanae, Novolazarevskaya and Syowa (Fig. 1) to investigate warm periods in the last decades (1957 AD–present and for the Neumayer Station 1981 AD–present). We expect Sanae and Novolazarevskaya to be most representative for the core due to their close proximity. A comparison of mean annual and mean summer (December–February) temperatures at the five stations reveals a strong temperature gradient between Sanae and Novolazarevskaya (7°C difference for the annual mean and 5.5°C difference for the summer months, 1960–91). The lowest long-term mean summer temperature was recorded at Sanae (-7.2°C) and the highest at Novolazarevskaya (-1.7°C). During the winter months (June–August) strong temperature gradients occur between Halley and Neumayer stations (about 4°C) as well as between Sanae and Novolazarevskaya (about 5°C difference). In the winter Halley has the lowest temperature (-28°C) and Novolazarevskaya the highest temperature in the area (-16.7°C). Note that the S100 core is located between Sanae and Novolazarevskaya stations where the horizontal air temperature gradient is strong.

We observe a large increase in the annual frequency of melt layers at depths corresponding to summers 1985/86 and 1986/87 AD. This finding agrees well with warmer summers reported by the meteorological stations located nearby the site (Fig. 5). All stations show positive summer temperature anomalies in 1986 while in 1987 AD only one station (Neumayer) has slightly negative summer temperature anomaly. Positive anomalies of summer air temperatures in the 1970s AD registered at all stations have their response in increased number of ice layers in the core (Fig. 5). The mean temperatures in summers 1960/61, 1961/62 and 1962/63 AD are below the average, according to the meteorological data, and there is only one ice layer found for each year. Years with three or more ice layers correspond to warmer than average summer temperatures observed at three or four meteorological stations (e.g. 1985/86 and 1986/87 AD). However, we also noticed years with several ice layers and no indication of a summer warmer than normal (e.g. 1967/68 AD).

Since the mean summer temperature might not necessarily indicate favourable melting conditions we used daily temperature data from the meteorological stations (courtesy of Dr Gert König-Langlo, AWI and Dr Gareth Marshall, BAS) to find a number of days with positive (> 0°C) air temperatures - Positive Degree Days (PPD) during summer months (December–February). However, we did not observe any straightforward connection between the melt layer frequencies as derived from the firn core and the number of days with positive temperatures. Summers 1990/91 and 1991/92 AD characterized high PDD, although there is no indication of significant melting during these years. Summers 1985/86 and 1986/87 AD, with four and three ice layers respectively, do not show an increased number with positive temperatures. As van den Broeke (2005) points out, the best indicator of melting would be amount of hours with positive air temperatures throughout the season. This kind of data is rarely available. It should also be noted that some daily data are missing and daily data for Sanae are not available.

Along the whole core, we found 20% years with two or more ice layers, whilst 120 years (46%) show no indication of melting; single-ice-layer years dominate in the last 100 years. We see this pattern on an annual scale as well as for decades and for 25-year periods (Fig. 6).

There are 14 summer seasons (summer peaks in $\delta^{18}O$)

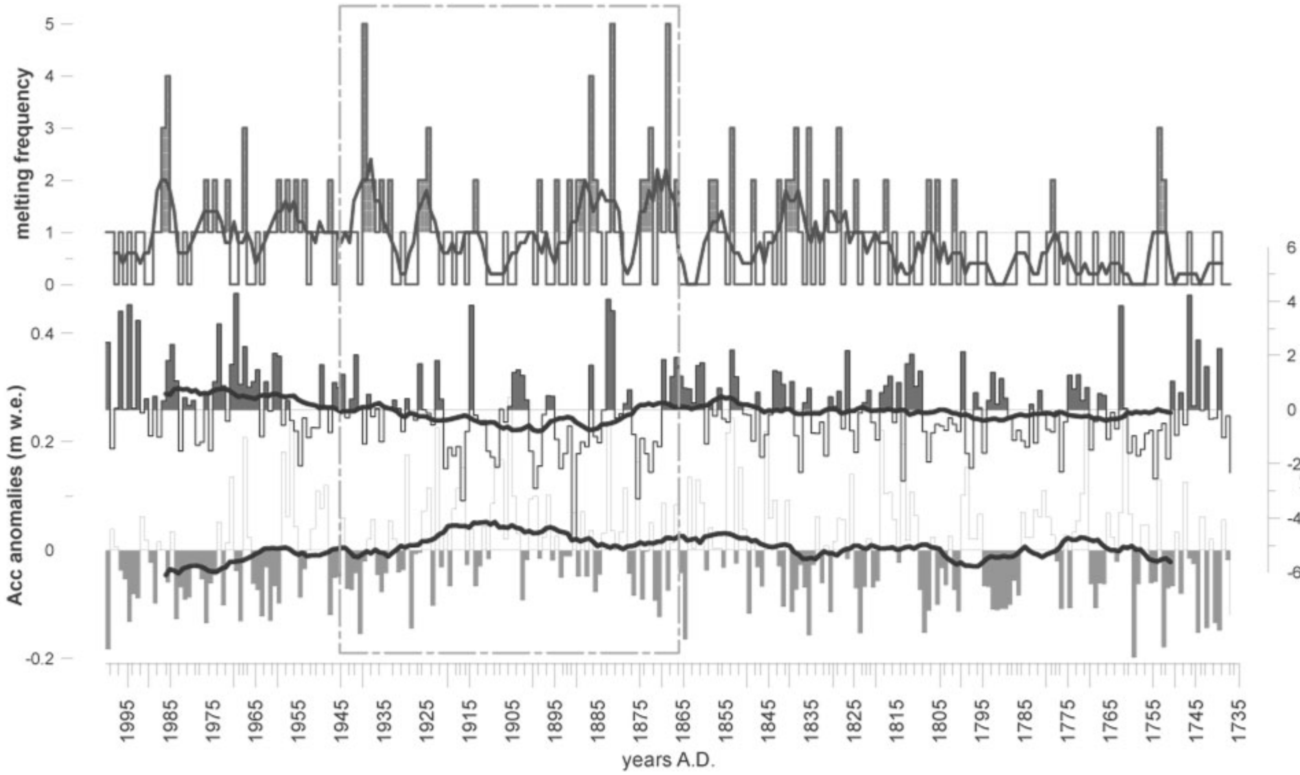

Fig. 7. Melt layer frequency (top) compared with annual anomalies in δ^{18}O (middle) and accumulation (bottom) records derived from the firn core S100. Thick curves in accumulation and δ^{18}O are 30-year running average and the thick curve in the melting frequency record is a 5-year running average.

with three or more melt layers. The oxygen isotope record indicates that in a few cases these did coincide with a positive annual oxygen isotope anomaly (e.g. 1882 AD with five ice layers and δ^{18}O anomaly of 3.7‰), while for several years the annual oxygen isotope anomalies could even be negative (e.g. 1940 AD with 5 ice layers but δ^{18}O anomaly is -1.3‰). The years 1882 AD and 1940 AD mark the beginning and end of a period where long-term accumulation and oxygen isotopes change. Around 1860 AD accumulation starts to increase, reaching its peak around 1900–20 AD, and decreases afterwards (Kaczmarska *et al.* 2004). Oxygen isotopes decrease between ~1860 AD and 1900–20 AD and then start to increase again. We noticed several years with high ice layer frequency at both ends of this time span while the section corresponding to 1890–1925 AD shows fewer ice layers (Fig. 7). We have not found any clear relation between the ice layer frequency, δ^{18}O and accumulation on short time scales (e.g. annual). As discussed recently (Helsen *et al.* 2005), oxygen isotopes may not be representative of temperature on annual to decadal time scales. On longer time scales, the frequency of melting events seems to correspond to both accumulation and oxygen isotopes for years 1890–1930 while there is no evident correlation for other periods (Fig. 7). These results indicate that melt layer frequency is difficult to explain with standard climate data and proxies derived from ice cores. Microclimate at the core location and the morphology of the local snow surface must play an important role as well.

Since precipitation events are common in the ice shelf zone of DML due to passing of cyclonic systems (van den Broeke *et al.* 2004a) and air temperatures can rise above 0°C during the summer months, we assume that some precipitation might be in the form of rain. Thus, both raindrops and meltwater can freeze within the snow pack creating ice layers. The position of the ice layers at depth, plotted against the δ^{18}O record, indicates that some ice layers actually belong to other than summer seasons (e.g. Fig. 3). We suspect such layers were created by meltwater that percolates further into the firn and refreezes there thus disturbing the seasonal pattern of melt events. Penetration of short wave radiation may produce sub-surface melting while the surface itself stays frozen as the result of outgoing long wave radiation (Liston *et al.* 1999). If there is snow precipitation in the summer, surface albedo increases and ablation is reduced. According to Dr Gert König-Langlo, rainfall occurs quite often at the Neumayer Station but the raindrops are frozen by the time they reach the surface. Summer melting also occurs regularly in the vicinity of Neumayer as well. The ice layers are formed usually at 50–100 cm depth and the melt water refreezes within the same year's strata (personal communication 2005).

Van den Broeke *et al.* (2004b) report high spatial and temporal variations in albedo and surface radiation balance for coastal DML in spite of the apparently homogeneous snow surface. The differences could be explained as a function of cloud cover, snow age and solar zenith angle, because clouds play a very important role in temporal albedo variations. One of the consequences of a strong correlation between albedo and cloud cover is that it partly offsets the warming effect that clouds usually have over highly reflective surfaces. Both clouds and precipitation events are frequent in the ice shelf zone of DML. Near the coast sublimation and melt cause extended periods of surface ablation in summer. Mass balance study based on the data from Automatic Weather Stations in DML (van den Broeke *et al.* 2004b) reveals that summer melting occurs regularly in the low-lying coastal part of DML (ice shelves). Melting shows large interannual variability and in summers with large events (e.g. 1997/98 and 1998/99) it may be even more significant than surface sublimation. Van den Broeke (2005) argues that the number of melt-hours provides a much more reliable indicator of melting than PDD. Since regional climate conditions are not well correlated with melt layer frequency, we believe that hourly air temperature data measured at a site are needed to provide reliable comparison.

Conclusions

Optical stratigraphy proved to be a useful tool to track melt layers in ice cores. We identified over 200 ice layers in a coastal ice core from DML for the last ~260 yrs (1737–2000 AD). However, our results indicate that melt layer frequency is difficult to explain with standard climate data and proxies derived from ice cores. Local microclimate and the morphology of the snow surface must play an important role as well. A relatively good agreement between the ice layer frequency and mean summer temperature recorded at the meteorological stations in coastal DML, Antarctica (Novolazarevskaya, Sanae, Neumayer, Syowa and Halley) can be seen for some years but not all. We have not found any straightforward connection between the melt layer frequency and Positive Degree Days. Difficulties with year-to-year correlation can be partly linked to the dating error in the core of ± 3 yrs. Percolating meltwater or even raindrops might refreeze at greater depths, disturbing the seasonal character of the ice layers. A major problem lies in the nature of melt layers. A single melt layer can be formed within several hours. Therefore, daily temperatures and number of hours with positive air temperatures, measured at a site, should be considered as better indicators of melt events.

We found no indication of increased melting in the last ~60 years despite an increasing trend in the $\delta^{18}O$ record. Melt event frequency varies considerably over time and periods with frequencies significantly different from the average are reported. Relatively few ice lenses at depths corresponding to approximately 1890–1930 AD agrees well with negative $\delta^{18}O$ anomalies and positive accumulation anomalies at that time. However, there is no significant correlation between climate proxies derived from the ice core and melt layer frequency for the entire period.

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