Inferences from stable water isotopes on the Holocene evolution of Scharffenbergbotnen blue-ice area, East Antarctica

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ABSTRACT. We show that it is possible to extract a high-resolution (annual) paleoclimate record from the surface of a blue-ice area (BIA). The variability of the surface stable-isotope values suggests that almost all the surface ice in Scharffenbergbotnen BIA, East Antarctica, is of Holocene age. The isotopic changes across the BIA show that the modern climate there is warmer than the climate in the early-Holocene optimum (11 kyr BP). A volume-conserving ice flow model for the BIA constrained by isotopic variability and layer thicknesses, and a series of ^{14}C ages indicate both that the BIA has been smaller than now, and that the surface velocities were considerably smaller during the Last Glacial Maximum. Changes in ice-sheet thickness drive the BIA towards present-day conditions. The relatively young age of the majority of the BIA also explains the lack of meteorite finds in this area, and may be typical for many BIAs in low-elevation nunatak areas.

1. INTRODUCTION

Antarctic blue-ice areas (BIAs) are known to have old ice at the surface (e.g. Whillans and Cassidy, 1983; Nishiizumi and others, 1989; Bintanja, 1999). Ablation in Antarctic blue-ice areas above 1000 m is overwhelmingly dominated by sublimation rather than melting (Bintanja, 1999). Such ice is likely to contain a high-resolution paleoclimate record that is easier to access than traditional deep ice cores. The dating of surface blue ice is, however, demanding. Previously, blue-ice samples from various BIAs have been dated by terrestrial ages of meteorites found on their surface (e.g. Whillans and Cassidy, 1983; Nishiizumi and others, 1989), by ^{14}C dating of ice (Van Roijen and others, 1995; Van der Kemp and others, 2002), by radiometric dating of tephra layers found at the surface of BIAs (Wilch and others, 1999) and by stratigraphic comparison with ice cores (Moore and others, 2006).

Isotopic composition of polar snow and ice has been regarded as a valuable temperature proxy in East Antarctica for decades (e.g. Lorius and Merlivat, 1977). Here, we make use of the ratios of heavy to light atoms of both oxygen and hydrogen expressed as δ^{18}O and δD values, respectively. The deuterium-excess, d ($d = (\delta D - 8)\delta^{18}O$), is assumed to depend mainly on the physical conditions in the source area for mid- and high-latitude precipitation. Changes in d are traditionally used as indicators of changes in the average temperature of oceanic moisture sources (Merlivat and Jouzel, 1979; Petit and others, 1991; Vimeux and others, 2001). However, Helsen and others (2006) showed that the vertical gradient in d excess over the moisture source area and the kinetic fractionation along the transport path have a prominent influence on the observed d values.

There are very few paleoclimate data records from Antarctic BIAs. The only continuous horizontal stable-isotope record, i.e. a δ^{18}O record extracted from an ice sample cut from the surface of a BIA along the flowline, has been extracted from Mount Moulton (76° S, 135° W; 2800 m a.s.l.) and covers 140 000 years (Popp and others, 2004). However, this record does not include the Holocene since that part of the BIA was covered by snow when sampling was done.

Here we focus on the surface blue ice in Scharffenbergbotnen BIA, Dronning Maud Land (DML), (74° S, 11° W; 1200 m a.s.l.), where there is some uncertainty in the dating. Some authors argue that glacial ice is present at the eastern end of the valley (Van Roijen, 1996; Grinsted and others, 2003), but others have suggested, in general terms, that BIAs in DML may have been an accumulation area during the glacial period (Bintanja, 1999). In this paper, we show that almost all the surface ice in the area is Holocene, based on the variability of the stable-isotope values. We study the spatial and temporal isotopic changes in the BIA in terms of climate variability, and compare results with other East Antarctic sites. Finally, we show results of a simple model on how the dynamics of the BIA may have evolved since the Last Glacial Maximum (LGM).

2. BACKGROUND

2.1. Study area

Scharffenbergbotnen (Fig. 1) is the best-studied Antarctic BIA. It is a valley located in the Heimefrontfjella mountain range at the edge of the Antarctic plateau about 350 km from the coast. Several studies have been made of its mass balance (Jonsson and Holmlund, 1990; Jonsson, 1992; Sinisalo and others, 2003), ice flow and surface age distribution (Van Roijen, 1996; Grinsted and others, 2003; Sinisalo and others, 2004), on the blue-ice surface properties (Bintanja and others, 2001) and on the moraines in the area (Lintinen and Nenonen, 1997; Hättestrand and Johansen, 2005).

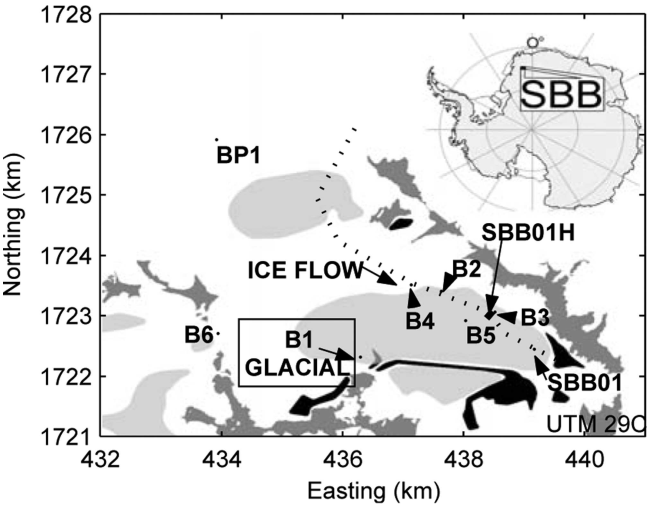

Fig. 1. Locations of the shallow blue-ice cores (B1–B5), firn core (B6), snow pit (BP1), the 100 m horizontal ice core (SBB01H) and the 52 m vertical core (SBB01) in Scharffenbergbotnen (SBB). Supraglacial moraines are marked in black, mountains in dark gray and blue ice in light gray. The ice flow is from northeast into the valley (dotted line). The ice passes through a small BIA and then terminates in the main BIA close to SBB01.

The meteorological conditions in the valley and surrounding area are described in detail by, for example, Bintanja and Van den Broeke (1995a, b), Bintanja (2000a, b), Bintanja and Reijmer (2001) and Reijmer (2001). The annual average temperature is about –20°C and wind speed is ~7 m s^{-1} (Reijmer, 2001). Scharffenbergbotnen is located in the lee side of the nunataks, and geostrophically and katabatically forced winds blow from easterly directions (Bintanja, 2000b). The precipitation is characterized by a highly intermittent accumulation record (Reijmer and Van den Broeke, 2003) with large spatial variations in the valley (Sinisalo and others, 2003). The present-day moisture source area is in the southern Atlantic Ocean (Reijmer, 2001; Helsen and others, 2006).

The meteorological conditions over the BIA differ from those over the snow-covered surroundings as the air over the BIA is warmer and the relative humidity is lower than over a snow site (Bintanja and Reijmer, 2001). These conditions contribute to the observed high sublimation rates of blue ice. Surface sublimation over the BIA is significantly higher than over snow (Bintanja and Reijmer, 2001), being >0.1 m w.e. a^{-1} at the southeastern end of the valley (e.g. Sinisalo and others, 2003). Slight surface melting occurs during a few high-insolation days in the BIA. The surface water film, however, is subsequently refrozen and removed by sublimation.

The main BIA in Scharffenbergbotnen is of the closed type, i.e. the ice has no outflow from the valley (Grinsted and others, 2003), and therefore it must have old ice at the surface if it is in steady state. According to geomorphological studies of Hättestrand and Johansen (2005), the difference between the surface elevation in Scharffenbergbotnen and outside the valley is greater today than when the ice sheet was thickest, which probably occurred during the LGM. The debris cover of the supraglacial moraines on the surrounding slopes in and outside Scharffenbergbotnen suggests that the ice surface in the valley was 200–250 m higher, and the elevation of the surrounding ice sheet only 50–150 m higher, at the LGM than today (Hättestrand and Johansen, 2005). The elevation decrease in the valley probably occurred gradually after the surrounding ice-sheet elevation had decreased after the LGM and ice overflow of the nunataks at the eastern end of the valley became insignificant. A decrease in surface elevation relative to the surrounding nunataks results in stronger katabatic flow, which has a positive feedback to the extent of a BIA (Van den Broeke and Bintanja, 1995). The moraine structures strongly suggest that the inner part of Scharffenbergbotnen must have been a local ablation area during the LGM; i.e. a BIA has long existed in the valley (Hättestrand and Johansen, 2005).

2.2. Sample locations

A 52 m long vertical ice core (SBB01 in Fig. 1) was drilled in the innermost part of the valley close to the end of the current flowline during the austral summer of 1997/98 (R. Bintanja and others, unpublished information). A 100 m horizontal ice core (SBB01H in Fig. 1) was collected, using electric chainsaws, from the surface of the BIA 1 km upstream from SBB01 in 2003/04. Approximately the top 20 cm was cut off from the samples in order to remove a possible refrozen meltwater layer in the high-insolation period, and to avoid any other disturbances from surface processes that may have influenced the ice composition.

In addition, a 10 m firn core (B6) and five 3 m shallow cores (B1–B5) were drilled in the austral summer 1999/2000 (Fig. 1). In the same field season, a 2 m snow pit (BP1) was also sampled at the northwestern entrance to the valley (Fig. 1). The details of the subsampling of the blue-ice cores and snow and firn samples are collated in Table 1.

2.3. Previous dating of Scharffenbergbotnen blue ice

Several blue-ice samples were dated using the ^{14}C method described by Van Roijen and others (1994) and Van der Kemp and others (2002) and converted to calendar ages using the radiocarbon calibration curve of Reimer and others (2004). The surface ages at the main BIA varied between 4000 and 14 000 years along the flowline (Fig. 1). These ages, however, have large uncertainties of up to several thousands of years. The ^{14}C age for the uppermost 45 m section of the SBB01 is 9300 ± 400 years (Van der Kemp and others, 2002) which corresponds to a calibrated calendar age of 10 500 (+700, –300) years. Unfortunately, a vertical age span cannot be determined for the ice core from the ^{14}C data.

Van Roijen (1996) used a numerical model of the ice flow in the valley based on the shallow-ice approximation and compared its results to the ^{14}C dating of the ice samples. He obtained surface ages of up to 60 000 years at the end of the flowline at the eastern end of the valley using three different surface velocity and mass-balance scenarios. Grinsted and others (2003) modelled the ice flow in the valley with a volume-conserving model which assumes constant ice-sheet geometry over time, i.e. steady-state flow. The flowline (Fig. 1) was chosen based on the measured velocity data (Van Roijen, 1996; Sinisalo and others, 2003) and is more realistic than the flowline that Van Roijen (1996) used, although the differences are not crucial. Grinsted and others (2003) used the measured accumulation and surface velocities (Van Roijen, 1996; Sinisalo and others, 2003) as input parameters, and obtained very old ages (~100 000 years) for the ice at the end of the flowline. The difference between the modelled ages is most likely due primarily to different grid resolutions at the end of the flowline where the ages are highest.

3. METHODS

3.1. Isotopic analysis

The $\delta^{18}O$ and δD analyses of the SBB01H and SBB01 cores were made at the Centre for Isotope Research, University of Groningen, The Netherlands. The $\delta^{18}O$ measurements were performed with a Sira-10 isotope-ratio mass spectrometer with an adjacent CO_2–H_2O isotopic equilibrium system. The δD measurements were performed using a continuous-flow system, consisting of a Eurovector chromium reduction oven coupled to a GVI Isoprime. The accuracy (combined uncertainty) of $\delta^{18}O$ analysis was ±0.06‰ and of δD ±0.7‰. The $\delta^{18}O$ analysis of the 3 m blue-ice cores, and the 10 m firn core and 2 m snow pit was performed at the University of Technology, Tallinn, Estonia, using a Finnigan-MAT Delta-E mass spectrometer. Combined uncertainty of the analyses was better than ±0.1‰. The $\delta^{18}O$ and δD are both presented with respect to the international consensus Vienna Standard Mean Ocean Water – Standard Light Antarctic Precipitation (V-SMOW–SLAP) scale (R. Gonfiantini, unpublished information). The accuracy of *d* excess is ±1.3‰.

3.2. Isotopic paleothermometer

We use the isotope record as an indicator of local temperature change in Scharffenbergbotnen and compare it with other sites from East Antarctica. Although the time-spans of the individual isotope samples from the blue-ice cores are not known, based on present-day accumulation rates (Sinisalo and others, 2003), it is plausible to assume that most of our samples span time periods of several years to centuries. Hence, the influence of seasonal extreme isotopic and temperature values that could invalidate the classical temperature interpretation of isotopic variability is minimized (Helsen and others, 2005). However, it is necessary to make corrections both for elevation changes in Scharffenbergbotnen during the Holocene and for different ocean surface isotopic composition in the early Holocene. Thus, we calculate a change in $\delta^{18}O$ values due to temperature change, $\Delta\delta^{18}O_{temp}$, as

$$\Delta\delta^{18}O_{temp} = \delta^{18}O_m - \left(\Delta\delta^{18}O_{EC} + \gamma_m \Delta\delta^{18}O_{SW}\right), \qquad (1)$$

where $\Delta\delta^{18}O_m$ is the difference between the average $\delta^{18}O$ values measured at two sites of different age (Fig. 1), $\Delta\delta^{18}O_{EC}$ is the change associated with elevation change in time, $\Delta\delta^{18}O_{SW}$ is the change in isotopic composition of ocean surface waters in time due to deglaciation and γ_m (= 0.6) is the temporal sensitivity of $\delta^{18}O$ to the changes in marine isotopic composition (Vimeux and others, 2002; Kavanaugh and Cuffey, 2003).

In addition, there are other factors, such as changes in the water-vapor source area (Kavanaugh and Cuffey, 2003), changes in precipitation seasonality (Werner and others, 2001) and changes in the strength of the temperature inversion (Van Lipzig and others, 2002), which may have influenced isotopic changes in the Holocene. We assume here that these factors are secondary and can be discarded. We justify this assumption for some cases in section 4.2.

The decrease in surface elevation of 200–250 m in Scharffenbergbotnen during the Holocene (Hättestrand and Johansen, 2005) corresponds to a change of 9.3–12‰ in δD (1.2–1.5‰ in $\delta^{18}O$) using the present-day altitudinal lapse rate for $\delta^{18}O$ values of 5.8‰ km^{-1} (Isaksson and Karlén, 1994). This lapse rate is calculated for $\delta^{18}O$ values measured from 10 m firn cores covering 15–30 years of accumulation along a traverse that crossed the Scharffenbergbotnen area. We calculate a standard error, σ_{EC}, for $\Delta\delta^{18}O_{EC}$ of ±0.1‰. The $\Delta\delta^{18}O_{SW}$ was about +1.1‰ at the LGM compared with the present value (Labeyrie and others, 1987), and it was still +0.2‰ at 10 000 years BP (Waelbroeck and others, 2002).

Table 1. Sampling depth/length, number of subsamples (*n*) and length of each subsample for the vertical blue-ice core B1–B5 and SBB01, for the firn core B6, the snow pit BP1 and the horizontal blue-ice core, SBB01H

Core	Depth/length m	Number of samples	Sample length cm
SBB01	25.9–31.1	25	1
		23	3
		10	18–26
SBB01H	100.4	87	45–140 (average 112)
	0.6	27	2
B1–B5	2.7–3.0	4	2
B6	10.0	28	2–24
BP1	2.1	17	2–3

The temperature change corresponding to a known $\Delta\delta^{18}O_{temp}$ can be calculated using the present-day spatial isotopic temperature gradient in Antarctica as a surrogate for the temporal isotopic temperature gradient (Delaygue and others, 2000; Masson and others, 2000; Jouzel and others, 2003). In this study, we use an isotopic temperature gradient of 1.16‰ K^{-1} from Isaksson and Karlén (1994). The gradient is greater than found elsewhere in Antarctica but it is calculated for samples drilled very close to our study area. We estimate that the error, $\sigma_{temp} = \pm 0.28$‰ K^{-1}.

4. RESULTS AND DISCUSSION

The mean values of the stable-isotope ratios, $\delta^{18}O$, and the population standard deviations (in ‰) for each core or pit are presented in Figure 2. The confidence interval (at 95% level) was less than ±0.7‰ for all the $\delta^{18}O$ mean values. Table 2 shows measured $\delta^{18}O$ and δD values and the population standard deviations (in ‰) for SBB01 and SBB01H.

4.1. Age estimation of blue ice

Different climatic periods have different signatures in stable isotopes (e.g. Petit and others, 1999). We determine whether the samples at a given site were deposited during a glacial or an interglacial period simply from the isotopic composition.

A rapid change of ~40‰ in δD (5‰ in $\delta^{18}O$) in Antarctic ice is an indicator of a change between interglacial and glacial climates (e.g. EPICA Community Members, 2006). Climate variability within the Holocene as measured along the EDML core (75° S, 0° E; 2900 m a.s.l.), the closest deep core to the study site in East Antarctica, causes changes of <2‰ in $\delta^{18}O$ in the centennial-scale variability, and the maximum difference in decadal means of $\delta^{18}O$ is ~5‰ for Holocene ice (H. Oerter, http://doi.pangaea.de/10.1594/PANGAEA.264634). The standard deviation of the $\delta^{18}O$ values measured from B2–B5 in Figure 1 is <1.8‰, and the difference between $\delta^{18}O$ values measured from B2–B5 and the present-day value of –28.5‰, taken as an average from B6 and BP1, is <4‰. In addition, geomorphological

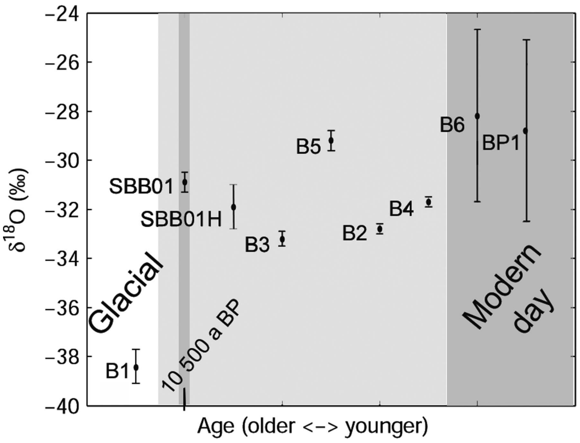

Fig. 2. Measured $\delta^{18}O$ values for the shallow blue-ice cores (B1–B5), firn core (B6), snow pit (BP1), the 100 m horizontal ice core (SBB01H) and the 52 m vertical core (SBB01) marked in Figure 1 with their population standard deviations. The samples are ordered by their relative age along the *x* axis from the sample with the oldest ^{14}C age (B1) to the firn and snow samples (B6 and BP1) representing the present-day values in the valley.

evidence suggests the blue-ice samples B2–B5 originate from a higher elevation (Hättestrand and Johansen, 2005). The correction of the elevation change would make the difference in $\delta^{18}O$ values between the blue-ice samples and present-day samples even smaller. Thus, we simply conclude that most of the main BIA in Scharffenbergbotnen is of Holocene origin.

SBB01, dated at 10 500 years, is located close to the bottom of the valley where the oldest surface ice along the current flowline occurs (Fig. 1). The most negative $\delta^{18}O$ value measured in the valley is, however, from B1. It is 9.9‰ lower than the present $\delta^{18}O$ value, which indicates that the ice in that particular sample, drilled from the southern margin of the main BIA (Fig. 1), originates from the glacial period. The oldest ^{14}C-dated sample was found in the same part of the BIA (Van Roijen, 1996), with a calibrated calendar age of more than 28 000 years BP.

In the high-resolution $\delta^{18}O$ data of a 60 cm long section from SBB01H we clearly see three annual cycles (Fig. 3). We determine, from the power spectrum of Figure 3, that the horizontal age gradient at that location is $\sim$5.4 years m^{-1}. The result agrees with the surface age gradient of 3–6 years m^{-1} determined by dating of internal radar reflection horizons close to the current blue-ice/snow transition zone along the flowline (Sinisalo and others, 2004). This was the only high-resolution section of the SBB01H. The flow model of Grinsted and others (2003) gives an almost constant surface age gradient over the BIA. It is therefore reasonable to extrapolate this age gradient over the 100 m horizontal ice core, SBB01H. Thus we find that the horizontal ice core covers about 540 years. Similarly extrapolating over the 1 km distance between SBB01 and SBB01H gives an age of about 5000 years for SBB01H, as the SBB01 core is dated at 10 500 years BP. This age is, of course, a rough approximation and we shall return to it later in relation to the flow model.

No significant periodicities were found in the high-resolution isotopic data from a 1 m section of the vertical core SBB01. We assume that the age–depth relationship is linear for the vertical ice core since the core penetrates only a small fraction of the total ice thickness (Herzfeld and Holmlund, 1990). The isochrones in the BIA, according to flow models, are strongly inclined at the SBB01 drilling site, which is close to the bottom of the valley where vertical flow dominates (Van Roijen, 1996; Grinsted and others, 2003). This means that the vertical core is not perpendicular to the isochrones and the annual layers seem much thicker since the core cuts them obliquely. As the ice is relatively old, we can expect it to have experienced more strain thinning of annual layers. We can also expect that diffusion will act to smooth high-frequency variability in the core, relative to the signals in SBB01H. Therefore it is not surprising that there are no high-frequency cycles present in the SBB01 core, and that the 5 m section of ice used to extract the mean isotopic values (Table 2) samples a large number of years.

4.2. Low-frequency changes

Several isotopic records from East Antarctica exhibit a clear early-Holocene optimum immediately following the end of the last ice age from 11 500 to 9000 years BP (Masson and others, 2000). Thus, it is plausible to assume that 10 500 year old SBB01 represents the early-Holocene optimum that is generally defined as the warmest climatic period during the Holocene. In Scharffenbergbotnen, however, our results show that the present-day climate is warmer than in the early-Holocene optimum. We use a value of –28.5‰ (average from B6 and BP1; Fig. 2) for present-day $\delta^{18}O$ in the valley. The change in $\delta^{18}O$ between SBB01 (Table 2) and the modern level is $\sim$2.4‰ (19‰ in δD). Equation (1) gives a $\Delta\delta^{18}O_{temp}$ value of 1.2 ± 0.2‰ for $\Delta\delta^{18}O_{SW} = 0.2 \pm 0.1$‰ and an elevation change of 225 m. According to the isotopic temperature gradient (Isaksson and Karlén, 1994), this corresponds to a warming of $\sim 1.0 \pm 0.3^{\circ}C$ since the early Holocene optimum.

In contrast to the measurements in Scharffenbergbotnen, Masson and others (2000) found an opposite change in several isotopic records in East Antarctica between the early-

Table 2. The mean values of the stable-isotope ratios $\delta^{18}O$ and δD and the population standard deviations (in ‰) for the vertical blue-ice core SBB01 and the horizontal blue-ice core SBB01H, the number of the samples and calculated deuterium excess, *d*

Core	Number of samples	Mean $\delta^{18}O$ ‰	Std dev. $\delta^{18}O$ ‰	Mean δD ‰	Std dev. δD ‰	Mean *d* ‰	Std dev. *d* ‰
SBB01H all	69[a], 87[b]	–31.9	0.9	-251.6	7.3	3.7	0.9
SBB01 all	58	–30.9	0.4	-238.5	3.0	8.6	1.1

[a]Number of $\delta^{18}O$ samples.
[b]Number of δD samples.

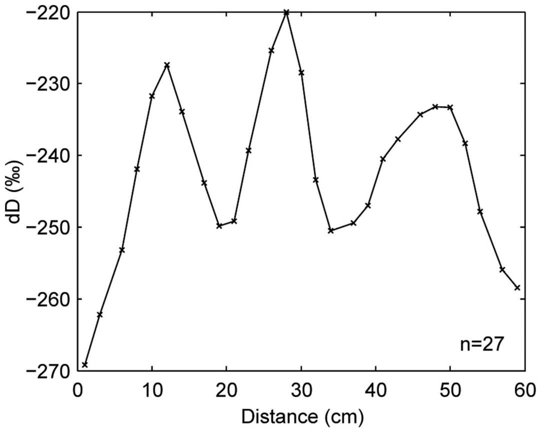

Fig. 3. Results of the high-resolution δD analysis measured from a 60 cm section of the horizontal core SBB01H.

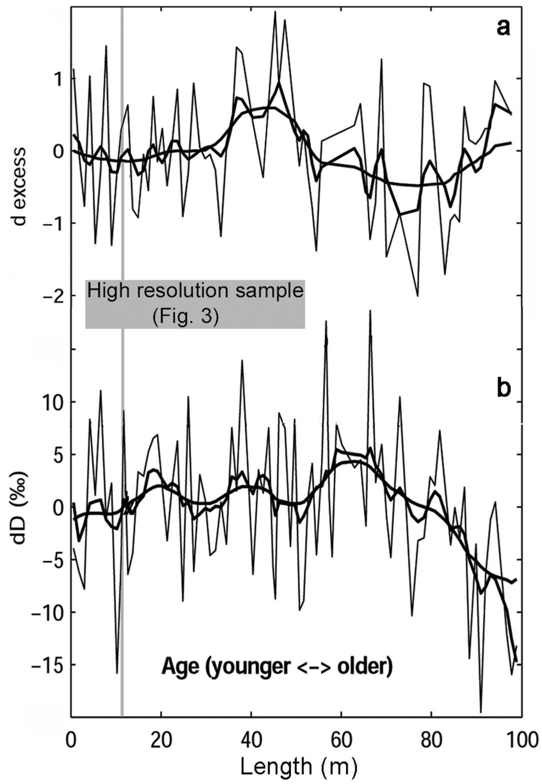

Fig. 4. The variability of *d* excess (a) and δD values (b) of the horizontal SBB01H core: the *d* excess and δD records (gray), long-term trend as the first reconstructed component of the singular spectrum analysis (SSA) (e.g. Ghil and others, 2002) using an embedding dimension of 10 (thick black line), and the partial reconstruction as the sum of the first and second component of the SSA (thick gray line). SBB01H is 100 m long and oriented along the flowline (Fig. 1). The youngest ice is found at $x = 0$. Older ice, ~540 years, is found downstream at $x = 100$ m.

Holocene optimum and modern levels. The decreasing trends found elsewhere in East Antarctica are probably the result of an overall Holocene increase in elevation of the East Antarctic ice sheet (Masson and others, 2000), due to increased Holocene accumulation rates (Ritz and others, 2001).

The SBB01 core has a 1‰ higher mean value in δ^{18}O (and ~13‰ higher δD) than the horizontal core SBB01H (Table 2). The δ^{18}O values of SBB01H are also lower than the present-day value of –28.5‰ by ~3.4‰ (27‰ lower for δD). We know that there was an elevation decrease of 200–250 m in Scharffenbergbotnen between the LGM and the present day (Hättestrand and Johansen, 2005), and that the elevation must have changed gradually. Thus, we use $\Delta\delta^{18}O_{EC} = 0.6‰$ and $\Delta\delta^{18}O_{SW} = 0$ for mid-Holocene and present-day values. From Equation (1) we find $\Delta\delta^{18}O_{temp} \approx -1.6 \pm 0.1‰$ between SBB01 and SBB01H, and $\Delta\delta^{18}O_{temp} \approx 2.8 \pm 0.2‰$ between SBB01H and the present-day samples. These changes correspond to a cooling of $\sim 1.4 \pm 0.4$°C and warming of $\sim 2.4 \pm 2.0$°C, respectively.

There is a decrease of 11‰ in δD (1.4‰ in δ^{18}O) in the last 40 m section at the downstream end of the SBB01H isotope profile (Fig. 4b). Oerter and others (2004) found that changes in precipitation seasonality in DML can cause trends in the δ^{18}O profile of ~2‰ within a 200 year period. That and influences of many source-region climate changes, however, are unlikely for the first half of the trend (60–80 m in Fig. 4) as they are expected to cause anticorrelated changes in *d* excess with δD (Kavanaugh and Cuffey, 2003; Oerter and others, 2004). Thus, using Equation (1) we calculate that the change of –4.6‰ in δD between 60 and 80 m (Fig. 4) corresponds to a temperature change of $\sim 0.5 \pm 0.2$°C using the temperature–isotope relationship of Isaksson and Karlén (1994).

4.3. Changes in blue-ice dynamics since LGM

In this paper, we have shown that the BIA has not been in steady state throughout the Holocene. However, according to the moraine studies (Hättestrand and Johansen, 2005), the inner part of Scharffenbergbotnen was a local ablation area at the LGM because otherwise the supraglacial debris would have been transported from the valley.

The generally young age of the surface ice is the result of the past mass-balance and flow regime. We can explore some possible scenarios with a volume-conserving flow model that does accommodate temporally variable surface velocity, ice thickness and mass balance along the flowline with parameterized variation of ice rheology with depth to produce particle trajectories and isochrones (Grinsted and others, 2003). There is no unique solution for how the BIA has changed over the last glacial cycle as there are only very few constraints on the surface age. We study three simple cases that produce surface ice ages comparable to those calculated from ^{14}C ages (Van Roijen, 1996; Van der Kemp and others, 2002). As the most accurate ^{14}C age was measured for SBB01, we define it to be the most important age to match. The cases are:

i. different surface velocity in the past;

ii. different accumulation rate in the past;

iii. a combination of cases i and ii.

In the following we discuss each case in turn.

i. Different horizontal ice velocity

There must have been less inflow through the northwestern gate to the valley (Fig. 1) at the LGM than today because the surface elevation difference between the valley and its surroundings was smaller. Additionally, there must have been inflow from other directions as the ice flowed over the mountains, at least at the eastern end of the valley (Hättestrand and Johansen, 2005), though there must have

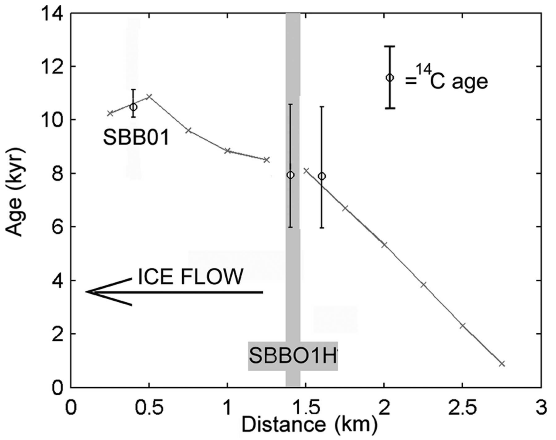

Fig. 5. The calibrated ^{14}C ages (Van Roijen, 1996; Van der Kemp and others, 2002) along the flowline (Fig. 1) and model output with a linearly changing temporal and spatial surface velocity and accumulation rate reaching the present-day values in 11 000 years. The starting accumulation rate was 0.13 m w.e. everywhere along the flowline and the surface velocity was zero. The error bars for SBB01 are calculated using a radiocarbon calibration curve of Reimer and others (2004). Only those of the other blue-ice cores (Van Roijen, 1996) that were located within 50 m of the flowline were plotted. There is thus an error associated with projecting the measurements onto the flowline. The relationship between error and distance was estimated by fitting a straight line to the relative difference between two ^{14}C measurements against their distance. The horizontal distance is measured starting from the bottom of the valley. SBB01 is located at $x = 400$ m and SBB01H at $x = 1400$ m.

been a net inward flow to preserve the BIA. Thus, the surface velocities must have been lower at the LGM than today.

It is not possible, however, to produce Holocene ages for SBB01 with the flow model using smaller surface velocities for the BIA in the past. On the contrary, the surface velocity would have to have been many times higher over the whole Holocene than the current measured velocity profile if it alone was responsible for the measured Holocene age. It is clear that different surface velocity alone cannot explain the young surface ice in the BIA.

The distance between SBB01 and the current equilibrium-line altitude, determined from the accumulation and ground-penetrating radar data, is ~2600 m (Sinisalo and others, 2004). Based on the geometry, the mean surface velocity needed for an age of 10 500 years BP for the SBB01 site is 0.5 m a^{-1}, if the surface velocities had been constant through time and the size of the BIA had not changed. This is 70% larger than the maximum velocity (0.3 m a^{-1}) that is measured in the valley (Sinisalo and others, 2003), and contradicts the evidence for lower surface velocities in the past. With a current average surface velocity of 0.14 m a^{-1} in the valley (Sinisalo and others, 2003), the 10 500 year old ice in SBB01 would have originated only 1500 m upstream. This is inside the present-day ablation area, so we conclude that the equilibrium line has probably moved over time and that the BIA was smaller in the past.

ii. Different accumulation rate

Many studies suggest increased accumulation in Antarctica during the Holocene in comparison with the LGM (e.g. Udisti and others, 2004). The results from the EDML core for the past 7000 years, however, show decreasing accumulation during the past 4000 years (Oerter and others, 2004). It is only possible to produce an age of 10 500 years for SBB01 with the flow model by increasing the accumulation rate earlier in the Holocene from the present observations. We get the best fit to the calibrated ^{14}C ages by adding a linear accumulation rate gradient of 2.2×10^{-5} m a^{-2} to the current measured accumulation rates at all positions along the flowline, so that the accumulation rates reach the present values in 11 000 years (Sinisalo and others, 2003). The model output gives a nearly linear surface age gradient over the whole BIA of about 4 years m^{-1}, which suggests SBB01H is ~6600 years old. The horizontal age gradient of 5.4 years m^{-1} estimated from the SBB01H high-resolution data (Fig. 3) is in reasonable agreement with the 4 years m^{-1} considering that only three cycles were measured isotopically, and natural accumulation variability over 3 years may typically be 30% (e.g. Isaksson and others, 1996; Sinisalo and others, 2003).

iii. Different ice flow regime in the valley

To model the scenario of both lower velocity and higher accumulation rate as suggested by the results of cases i and ii, we choose to linearly change the temporal and spatial surface velocity and accumulation rate for the flow model. We assume that the whole valley was an accumulation area in the glacial period (prior to 11 000 years BP) with an accumulation rate of 0.13 m w.e. everywhere along the flowline, and a starting velocity of zero. We let the surface velocity and the accumulation rate change linearly over time so they reach the present values at 0 years BP. This leads to an ablation area, i.e. a BIA, with surface ages matching the ^{14}C ages, even if the whole valley begins as an accumulation area and there is no inflow through the northwestern gate (Fig. 5). Of course, this scenario is not modelled realistically as the flow model is purely prescriptive, but it is included here to suggest the possibility of negligible ablation area in the last glacial period.

The best-fit model to the ^{14}C ages gives a surface age gradient of ~2.8 years m^{-1} between SBB01 and SBB01H. This suggests that SBB01H is about 8000 years old (Fig. 5). In general, the modelled surface age gradients agree with the earlier studies of dated GPR reflection horizons that gave values of 3–6 years m^{-1} at the firn/blue-ice transition zone (Sinisalo and others, 2004).

5. CONCLUSIONS

In this study we show that most of the main BIA in Scharffenbergbotnen is Holocene ice, based on the δ^{18}O values in blue ice and snow. The δ^{18}O values in SBB01 support the previous ^{14}C dating of SBB01 and rule out the possibility that the ice close to the bottom of the valley originates from the East Antarctic plateau or from a glacial period.

The oldest surface ice in the valley was found close to the moraines on the southern side of the main BIA in Scharffenbergbotnen where the δ^{18}O value was most negative (sample B1 in Fig. 1). The calibrated calendar age (Reimer and others, 2004) at that part gave an age >28 000 years BP (cf. Van Roijen, 1995: ^{14}C age >24 000 years BP). The oldest ice may have remained at the southern margin 'isolated' from the main flow. However, there is no indication of where this ice originates.

We showed that it is possible to extract a high-resolution paleoclimate record from the BIA even with an annual resolution. However, we need a longer horizontal isotopic profile from the BIA in order to study how the surface age gradient varies and to determine the age of SBB01H reliably.

The differences in stable-isotope values between blue-ice and firn samples imply that the modern climate is about $1.0 \pm 0.3°C$ warmer than the climate in the early-Holocene optimum in Scharffenbergbotnen. Further, the 10 500 year old SBB01 originates from a warmer period than the mid-Holocene SBB01H.

According to our simple flow modelling it is possible that the whole of Scharffenbergbotnen was an accumulation area at the LGM. However, previous studies of supraglacial moraines and ^{14}C dating, together with $\delta^{18}O$ values at the southern margin of the main BIA, indicate that the BIA existed during the LGM. Therefore we suggest that the BIA was smaller than it currently is, and that the surface velocities were considerably smaller at the LGM. The young age of the major part of the BIA also explains the lack of meteorite finds in this area, and may be typical for many BIAs in low-elevation nunatak areas, where the ice-sheet elevation changes at the glacial termination are likely to have been most pronounced (Pattyn and Decleir, 1998).

It is clear that the evolution of the BIA requires a full diagnostic flow model, and we are presently setting up a finite-element scheme solving the full polythermal Stokes equations (Le Meur and others, 2004).

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