Antarctic blue ice areas (BIAs) - towards extracting paleoclimate information

Anna Sinisalo[1]* and John C. Moore[1,2,3]
[1]Arctic Centre, P.O.Box 122, 96101 Rovaniemi, Finland
[2]Thule Institute, University of Oulu, Finland
[3] College of Global Change and Earth System Science, Beijing Normal University, Beijing, China
*Current affiliation: University of Oslo, Department of Geosciences, Norway
Corresponding author: a.k.sinisalo@geo.uio.no

Abstract
We review the current scientific knowledge about Antarctic Blue Ice Areas (BIAs) with emphasis on their application for paleoclimate studies. Substantial steps have been made since the review by Bintanja (1999), in particular the question of dating the archive of ancient ice found on the surface of BIAs has progressed with advances in ^{14}C measurements, tephrachronology, and geomorphological evidence giving better constraints to more sophisticated ice flow models. Flow modelling also provides information about past changes in ice flow velocities, accumulation rates and ice sheet elevation. The availability of gas composition in vertical cores from BIAs allows matching to the well-dated global records of greenhouse gas variability over the last glacial-interglacial cycle and longer. It is clear from the limited number of studies to date that BIAs from different regions have quite different histories of formation and preservation, and that they are intimately linked to the response of their surrounding ice sheets to climate variability on glacial-interglacial time-scales. Looking to the future, climate records from BIAs are expected to provide information on variations in Southern Ocean processes as well as ice sheet evolution within the East Antarctic ice sheet at the thermal transition from cold based to warm based ice.

Keywords: Ice flow modelling, radar, ice cores, dating methods

1 Introduction

Antarctic paleoclimate data are available not only from deep ice cores but also from the surface of Antarctic blue ice areas (BIAs). Many of these BIAs are known to have very old ice at the surface (Whillans & Cassidy 1983, Bintanja 1999). The surface blue ice can also provide higher temporal resolution than usual in deep ice cores (Moore *et al.* 2006, Sinisalo *et al.* 2007), and BIAs are often located in coastal and mountainous areas where deep cores are not currently available. Thus, a series of shallow cores or samples from a horizontal surface transects from BIAs may complement the deep ice core insight of climate history (Sinisalo *et al.* 2007, Moore *et al.* 2006, Custer 2006, Popp *et al.* 2004).

The first scientific description of an Antarctic BIA was given by Schytt (1961) though there is a good general introduction by Giaever (1954) concerning the first expedition to visit a BIA. The expedition members discovered the amazing stability of the snow-blue ice boundary by placing match sticks at the boundary –many of which were still in place and not buried several months later. Schytt (1961) noted that crystal size of the blue ice indicated a deep source for the ice. Crary and Wilson (1961) first described the surface characteristics of BIA and discussed formation of BIAs by horizontal compressive forces with katabatic winds removing snow accumulation. Much of the early scientific interest in Antarctic BIAs was due to their nature as meteorite collectors (Cassidy *et al.* 1977, Whillans & Cassidy 1983, Nishiizumi *et* *al.*1989). The initial discovery of meteorites was made in the South Yamato BIA (Fig. 1) by Yoshida and others (1971) in 1969. Since then, more than 25 000 meteorites have been found on Antarctic BIAs (Harvey 2003). Consequently, the ice flow regime was studied and ice flow models were published for several BIAs to explain the meteorite findings (Naruse & Hashimoto 1982, Whillans & Cassidy 1983, Azuma *et al.* 1985). When researchers started to understand the flow regime of the BIAs, they got also interested in the paleoclimate record stored on the surface of BIAs (Bintanja 1999, Moore *et al.* 2006, Sinisalo *et al.* 2007, Custer 2006, Popp *et al.* 2004, Corti *et al.* 2008)

Bintanja (1999) stated that the easily recoverable ancient surface ice of the BIAs could be of great value for paleoclimatic purposes if the dynamics and the internal structure of the Antarctic BIAs were better known. In principle, a simple trench in the surface along a flow line of a BIA would yield ice samples of the full range of ages from present day at the upstream snowline to ice that can be as old as the particular BIA. However, the dating of the surface blue ice is demanding since the traditional ice core dating methods cannot be easily applied for the BIAs where the ice flow history is often complex and layering not horizontal. Blue ice samples from various BIAs have been dated using the terrestrial ages of meteorites found on their surface (e.g. Nishiizumi *et al.* 1989, Whillans & Cassidy 1983), by ^{14}C dating of air trapped in the ice (van Roijen *et al.* 1995, van der Kemp *et al.* 2002), by radiometric dating of tephra layers found at the surface of BIAs (Wilch *et al.* 1999), by stratigraphic comparison with well-dated ice cores (Moore *et al.* 2006), and by combination of several methods (Sinisalo *et al.* 2007). There are several BIAs such as South Yamato and Allan Hills (Fig. 1) where the surface blue ice is estimated to be tens or hundreds of thousands of years old (Nishiizumi *et al.* 1989, Whillans & Cassidy 1983, Welten *et al.* 2000), and the oldest terrestrial ages of meteorites are about 2 millions years. However, there are still only a few paleoclimate records from BIAs although potentially both the record length and temporal resolution may be higher than those from any of the other Antarctic ice cores.

Bintanja (1999) listed many open questions related to the BIAs, for example their general stability under different climate regimes, and the surface age gradient of individual BIAs of interest. In this paper, we review the latest knowledge of the BIAs emphasising paleoclimate applications.

2 Antarctic blue ice areas (BIAs)

BIAs are bare ice fields (Fig.2) that the wind keeps clean of snow, and their area can vary from a few hectares, e.g. many small BIAs in coastal mountainous regions of East Antarctica, to thousands of square kilometres, e.g. South Yamato BIA. Blue ice extent may vary because of weather, seasonal effects, and climate change. These changes, however, have not been well quantified.

Bintanja (1999) defines BIAs as areas where surface mass balance is negative, sublimation forms the main ablation process, and surface albedo is relatively low. This definition of BIAs specifically excludes melt-induced bare ice areas which are located on slopes in coastal areas where climate conditions together with favourable surface orientation favour surface and near-surface seasonal melt. These areas were identified by their spectral reflectance signature in satellite imagery by Winther *et al.* (2001) who divided observed bare ice regions to melt-induced and wind-induced BIAs. While melt-induced bare ice represents a minority of

Antarctic blue ice they are significant in that they are used as water supplies to Antarctic bases. For example, the Finnish and Swedish bases Aboa and Wasa use a melt area at Basen nunatak, East Antarctica. Melt-induced bare ice fields are also used as landing strips for wheeled aircraft such as the runaway close to the Novolazarevskaya station in Schirmacher Oasis, East Antarctica. For many Antarctic visitors these bare ice regions are their only experience of blue ice, and this often leads to false generalisation about Antarctic BIAs. According to the definition sublimation is the dominant ablation mechanism of a BIA, and it is clear that only the wind-induced BIAs can be useful for paleoclimate studies.

2.1 Geographical distribution

BIAs cover about 1 % of Antarctic surface area (Bintanja 1999). They are scattered widely over the continent appearing mainly in the vicinity of mountain ranges and nunataks as they are likely to form at locations where ice flow is dammed by outcropping nunataks or slowed down by sub glacial bedrock ridges (e.g. Faure & Buchanan 1991), precipitation is low and the annual mean wind speed relatively high (van den Broeke & Bintanja 1995). These phenomena explain why Antarctica is the only place on Earth where BIAs rather than simple melt areas exist.

Only a few BIAs have been studied in detail. They are located (see Fig. 1) in the vicinity of South Yamato Mountains (e.g. Naruse & Hashimoto 1982, Azuma *et al.* 1985, Moore *et al.* 2006), the Allan Hills (e.g. Faure & Buchanan 1991, Schultz *et al.* 1990, Spikes 2000), Mount Moulton (e.g. Wilch *et al.* 1999, Dunbar *et al.* 2008), and Heimefrontfjella mountain range in Dronning Maud Land (e.g. Bintanja 1999, Sinisalo 2007). In addition, there are other BIAs, such as Frontier Mountain BIA (Perchiazzi *et al.* 1999) that have been studied less intensively (Fig. 1), and many have simply been utilized as meteorite stranding grounds (e.g. Welten *et al.* 2000). Paleoclimate data have been collected from South Yamato (Nakawo *et al.* 1988, Machida *et al.* 1996, Moore *et al.* 2006), Mount Moulton (Custer 2006, Popp *et al.* 2004), and Scharffenbergbotnen in the Heimefrontfjella mountain range (Sinisalo *et al.* 2007).

2.2 Ice flow

The definition of the BIAs leads to a net upward component in the ice flow pattern in those areas. Thus, old ice layers originally buried deep in firn flow up to the surface and consequently there is old ice at the surface of the BIA (Fig. 3).

The ice flow regime, however, tends to be complex at BIAs where bedrock topography and outcropping nunataks mix the flow. This is the case even for large BIAs that extend up well upstream of surface outcrops such as South Yamato mountain BIA (e.g. Moore *et al.* 2006) and on smaller scales in Allan Hills (Grinsted *et al.* 2003). On the other hand, present-day flow regimes may be simple e.g. Scharffenbergbotnen (Grinsted *et al.* 2003, Sinisalo *et al.* 2004, 2007). However, the flow history may also have varied significantly over a glacial/interglacial cycle which makes dating of surface ice more complicated. The surface ice used for paleoclimate purposes to date varies from Holocene ice (Sinisalo *et al.* 2007, Sinisalo 2007) up to about 150 000 years (Popp *et al.* 2004).

2.3 Meteorology

The formation and maintenance of BIAs is facilitated by specific meteorological conditions induced by nearby mountains (Bintanja & van den Broeke 1995a, 1995b). According to Bintanja & Reijmer (2001), the meteorological conditions over the BIA differ from those over the snow covered surroundings due to

1) differences in surface characteristics, such as albedo, extinction characteristics for solar radiation, and surface aerodynamic roughness, between blue ice and snow, and
2) differences in topographic setting or nearby orography.

The air over the BIA is warmer and the relative humidity is lower than over a corresponding snow site, and these conditions contribute to the significantly higher sublimation rates over BIAs than over snow (Bintanja & Reijmer 2001). In general and depending on local topography, this can be attributed to adiabatic warming of large-scale katabatic flows (such as descending over much of East Antarctica), aided by diabatic warming due to radiation, and by turbulent mixing of warm air from aloft into the boundary layer (Bintanja & Reijmer 2001). The surface winds over BIAs are generally gustier than those over adjacent snow fields but the average wind speeds are comparable with each other (Bintanja 1999).

2.4 Surface characteristics

Antarctic BIAs have special surface characteristics such as low albedo, and aerodynamic smoothness in comparison to the surrounding snow and ice covered areas (Bintanja 1999). These characteristics aid formation and maintenance of the BIAs (Bintanja & van den Broeke 1995a, 1995b, Bintanja & Reijmer 2001). The surface of BIAs is generally rippled (Fig. 4). Bintanja *et al.* (2001) suggest that the only possible mechanism for blue-ice ripple formation is sublimation, occurring whenever there is wind forcing. The orientation of the crests of the ripples is perpendicular to that of the direction of the strongest winds. The measured wave-heights in different BIAs vary between 2 and 10 cm, and the wave-length between 5 and 24 cm (Bintanja *et al.* 2001, Mellor & Swithinbank 1989, Weller 1968). The annual average-wave height does not vary over time. The wave-height increases in the summer as the troughs of the ripples experience more sublimation than the crests, but this increase is compensated in winter (Bintanja *et al.* 2001). On the other hand, if the seasonal temperatures approach within a few degrees of melting the peaks are smoothed and melt can partially fill the troughs.

Dust or tephra bands (Fig. 4) are found in the surface of many BIAs where they usually appear perpendicular to the ice flow (Bintanja 1999) althought the dust bands in the Allan Hills and the Yamato Mountains are reported to curl back as much as 180° in some cases (Koeberl 1990). It is often extremely disorientating on BIA, however, and the local ice flow directions found by measurement can be quite different from intuitive ideas, and in our experience the dust bands always appear to be perpendicular to flow, provided that large gradients in ablation rates or large surface elevation changes (relative to the dip angle of dust layers in the ice) are not present.

The origin of the visible layers is of great interest. Dust bands in Allan Hills (Nishio *et al.* 1985), Mount Moulton (Dunbar *et al.* 2008, Wilch *et al.* 1999), South Yamato (Nishio *et al.* 1985), Frontier Mountain (Perchiazzi *et al.* 1999) and Lewis Cliff Tongue (Koeberl *et al.* 1988) have been shown to be of volcanic origin and they may be correlated with individual Cenozoic volcanoes in Antarctica and sub-Antarctic regions (Koeberl 1990). When volcanic eruptions disperse large quantities of volcanic dust over Antarctica, falls on snow accumulation areas, it will be buried under new snow layers and incorporated in the ice, then transported to the ablation zones with the ice flow. Such dust bands on the surface BIAs constitute isochronous layers.

There are also other possible sources for soluble and insoluble impurities than volcanoes. The most important other sources according to Koeberl (1990) include

1) material from sub glacial bedrock debris scraped up from the ground by the movement of the glacier. In practice this may be very difficult to find given the usual presence of moraines near the mountain terminus of a BIA where the bedrock may accumulate.
2) cosmic particles falling as micrometeorites or meteorite ablation spherules. These are known to constitute a sizable fraction of dust in central east Antarctica regions such as Dome Fuji (Nakamura *et al.* 1999), and so may be found in the down-slope flow line outcrop BIA of South Yamato Mountains; and
3) continental and marine dust and aerosols transported by wind. These probably constitute much of the material in the BIA of Dronning Maud Land which is far from active volcanic sources.

There are also other possibilities for the origin of the dust such as either surface or bottom crevassing. Crevasse traces, however, can be rejected on several grounds:

i) ablation rates are generally 5-10 cm a^{-1} on BIAs (Bintanja 1999, Sinisalo *et al.* 2003) which means that even deep crevasses (50 m deep) survive only about 1000 years, and clearly many bands are far older than that;
ii) bands are usually hundreds of metres long and often very smooth, which is certainly unlike most crevasse traces;
iii) direct investigation of crevasses in accumulation areas shows no evidence of dark material accumulating within them.

Other possible sources of banding considered are changes in flow of the ice due to surrounding ice sheet elevation changes dragging in moraine from other inflows, organized cryoconite separation similar to frost-heaving produced stone polygons and stripes, and ogive-like formation due to seasonal changes in flow over an icefall. However, all the mechanisms while possibly explaining some specific bands, cannot explain the general widespread banding on virtually every significant BIA we have observed. The widespread occurrence of bands, and the observation that only fine grain material is present in them below about ten cm depth also suggest that the bands were formed on the snow surface far enough away from the nunataks. This would mean perhaps up to tens of kilometres from the local nunataks. Hence, it can be assumed that the bands are isochrones and can therefore be reliably used as indicators of ice flow and ablation. There presence also has implications for general climate conditions that allowed widespread dust covering of the snow surface in the BIA accumulation areas.

Cryoconite holes can be found on the surface of low altitude BIAs (Bintanja 1999). They indicate melting, which may be purely internal due to solid-state greenhouse effects. The holes are round, bubble-free patches of ice (Fig. 4) that form due to absorption of solar radiation by dark particles or stones, causing their temperature to rise above the melting point (Bintanja 1999). Small stones and dirt will slowly sink into the ice until a depth where shortwave heating is diminished and equals conductive cooling.

Cryoconite holes are often preferentially formed on the bands on the ice surface. This is most likely due to the slightly lower surface of the band caused by its higher ablation rate allowing windblown rock particles to settle on the ice surface. The coarser cryoconites are confined in the upper decimetre, with only fine grain material beneath it. The abundance of cryoconites can be so significant that it influences the ablation rate of the glacier surface (Takeuchi 2002) due to its dark colouration and capacity to absorb solar radiation (Fig. 4). This solid-state greenhouse effect produces a favourable, watery niche for photosynthesis and, thereafter, bacterial respiration and grazing. Here viruses, bacteria (especially cyanobacteria), algae and fungi are typically dominant and they combine to form a soil-like granule (called cryoconite). These biota include organisms up to 10^5 years old that can be resuscitated and then thrive alongside younger organisms on the ice surface in a so-called *supraglacial ecosystem* (Hodson *et al*. 2005).

The depth of cryoconite holes varies seasonally. In winter after wind ablation and no solar heating, the dark material is closer to the surface than after the summer when the darker material has melted deeper into the ice. On Scharffenbergbotnen, for example, some debris from an old weather station destroyed by high winds can be seen within a few centimetres of the surface before summer begins in November, but by end of January the aluminium parts are 5-8 cm beneath the ice surface. Ablation in cryoconite holes can be a significant practical problem if the bare ice is used as a landing strip for wheeled aircraft. In an effort to improve the ice runway at Troll station, thousands of cryoconite holes were individually cleaned by removing all the cryoconite material from the melt water in the holes. It is not known how frequently this procedure needs to be repeated as in the long term as winds from the surrounding nunataks will surely provide more cryoconite raw material.

Snow patches often occur on BIA. While snow is often blown off a BIA, it can persist on the surface for many weeks to several years. The strong radiation, gusty winds and thermal gradients near the surface ice mean that the snow is quickly transformed into hard firn that becomes quite securely attached to the ice surface. However, the violent storms that can sweep BIA seem to be capable of removing even such dense firn from ice surfaces and preserving an ice surface that may be often hidden beneath up to 50 cm of hard-packed snow.

2.5 Classification of blue ice areas

BIAs can be divided according to their flow regime (Fig. 3, Grinsted *et al.* 2003) into

1) open type of BIAs, and
2) closed type of BIAs.

This division is relevant to the surface-age distribution of the BIA and hence the paleoclimate record that can be produced. The ice flow in an open BIA is not dammed by mountains, nunataks or bedrock topography but ice flows through the BIA and the oldest layers of ice do not come to the ice surface at all. The dipping angles of the outcropping isochrones are smaller than at the surface of the closed BIA. Closed BIAs are located at mountain ranges where their flow is dammed and the oldest layers of ice will be found in the surface closest to the mountains as if the layers had climbed up the mountain slope (Fig. 3). The South Yamato BIA is the largest open BIA, while e.g. Scharffenbergbotnen a closed BIA. Locally flow pattern in large open BIA can trap flow in cul-de-sacs and these are where the oldest terrestrial ages of meteorites have been found, particularly in Manhaul Bay in the Allan Hills BIA.

Takahashi *et al.* (1992) divided BIAs into four types that were based on the geographical setting and ice flow characteristics of the BIA. This geographical classification of BIAs was reproduced by Bintanja (1999), and it can be useful in understanding the formation and maintenance of a BIA:

1) Type I BIAs are associated with mountains protruding through the ice. They are situated in the lee of an obstacle, which acts as a barrier for snowdrift. The length of these BIAs can be estimated to be roughly 50 to 100 times the height of the obstacle relative to the ice surface (Takahashi *et al.* 1992). This is the most common type of BIA (Bintanja 1999). Scharffenbergbotnen is an example of the type I BIAs
2) Type II BIAs are located on a valley glacier. The descending katabatic winds cause net erosion of the surface and a local divergence of snowdrift, eventually leaving bare blue ice.
3) Type III BIAs are located on relatively steep slopes without mountains protruding through the ice. The increasing surface slope accelerates the down slope katabatic winds, causing a divergence of snowdrift transport similar to type II (Takahashi *et al.* 1992). The South Yamato mountains BIA is of this type.
4) Type IV BIAs are situated at the lowest part of a glacier basin. Accelerating katabatic winds in the basin remove the snow from the surface.

3 Variability of blue ice extent

The relations between blue ice extent and climate are not straightforward. Blue ice extent is sensitive to various climatic parameters, and climate change will affect the processes creating blue ice in several ways. Changes in air temperature will affect energy available for surface ablation, and changes in precipitation, and wind direction and strength will affect accumulation. (Orheim & Lucchitta 1990)

It is clear that large reductions in exposed BIA are more likely than large increases, and that increasing a BIA can be expected to take a longer time than decreasing it (e.g. Bintanja 1999, Brown & Scambos 2004). As a BIA cannot expand onto a nunatak area, aerial increase must take place into adjacent snow fields by the relatively slow processes of either dry snow metamorphosis, or exposure of sub-surface ice (Orheim & Lucchitta 1990). A decrease of the BIA, on the other hand, can happen quickly as a result of increased accumulation or possibly of changed wind patterns (Bintanja & van den Broeke 1995a). Seasonal and even annual variations in BIA extent due to snow accumulation events may be large and significant area reductions may occur (Brown & Scambos 2004). Minimum in the BIA extent is reached in winter (Brown & Scambos 2004). Snow may accumulate over a long time period but then rapidly removed by a large storm in a few days. This was observed in Scharffenbergbotnen (Fig. 1) where a severe storm in January 2007 removed old firn and surface snow up to 50 cm thick from the underlying ice. There have been attempts to estimate areal changes in individual BIAs (Brown & Scambos 2004, Spikes 2000, Orheim & Lucchitta 1990), and the total extent of Antarctic BIAs using satellite images (Winther *et al.* 2001). Increasing accumulation leading to a general decrease in blue ice can be easily detected, whereas increased ablation is more difficult to observe and requires more permanent change before it will be noticed. Since the possible permanent changes in blue ice extent are very slow and satellite images exist only over some decades, no change has been observed that could be interpreted as a climate signal.

When a BIA has formed, it tends to persist due to two conservative feedback processes (Bintanja 1999). These processes may enable BIAs to persist and possibly expand in the downwind direction. Firstly, the relatively low albedo of the blue ice increases the absorption of solar radiation, which increases the energy available for sublimation. Secondly, the smooth blue ice surface prevents drifting snow, and even quite thick firn and snow pack from becoming permanently attached resulting in zero accumulation in the longer term. All measurements of ablation rates on BIAs above 2000 m are remarkably similar (Bintanja 1999), which implies that changing climate does not affect ablation rates of high elevation BIAs greatly, and that there is a negative feedback between BIA extent and climate.

Although BIAs seem to be rather stable under different climate regimes due to the feedback mechanisms described above, climate change also affects ice flow patterns and may even reverse the ice flow direction in some areas. This can have a significant influence on the BIA extent over the long term as the specific ice flow field of the BIA will determine how stable a BIA is over glacial-interglacial periods (Bintanja 1999, Coren *et al.* 2003, Sinisalo *et al.* 2007). Allan Hills, for example, is the location of the oldest know terrestrial age meteorites discovered. However, the existence of very old ice does not necessarily imply a continuous temporal climatic record. Coren *et al.* (2003) suggest that the general ice flow has partially changed its direction in glacial-interglacial cycles in Allan Hills, and the ice with metorites travel in a flip-flop mode across the area. As the ice sheet is thinner in warm periods, open-type BIAs form on the lee side of bedrock rises, and some of the ice may be drawn then back to these ablations areas allowing meteorite concentrations to build up. In cold periods the regional ice flow is likely to be reduced but the flow is more uniform as the ice sheet is thicker (Coren *et al.* 2003).

It is obvious that the changes in ice sheet thickness control the existence of the BIAs in the mountainous areas (Bintanja 1999, Sinisalo 2007). Bintanja (1999) suggested that many of the type I BIAs currently existing in Dronning Maud Land did not exist during at LGM. He found that the height of the outcropping mountain in comparison to the ice surface height seems to be a critical variable in the temporal variations of type I BIA extent, as a thickening of the ice sheet leading to submergence of a nunatak will tend to make the BIA disappear. The ice sheet elevation changes at the glacial termination are likely to have been most pronounced in the nunatak areas a few hundred kilometres from the coast (Pattyn & Decleir 1998). The surface of the major part of the East Antarctic plateau ice sheet may have been about 100 m lower (Ritz *et al.* 2001, Pattyn 1999, Jouzel *et al.* 1989), but in transition areas from cold-based to warm-based ice flow at the margins of the plateau, hundreds of meters higher in the last glacial than at present (Näslund *et al.* 2000, Hättestrand & Johansen 2005). Surface elevation in Scharffenberbotnen, for example, used to be 200-250 m higher at LGM when surrounding ice sheet flew in to the valley over its side walls (Hätterstrand & Johansen 2005). The surface elevation started to decrease when lowering of the surrounding ice sheet reached a point where the outcropping mountains made a barrier to ice inflow over the side walls. The decrease of the surrounding ice sheet elevation enabled the BIA in the valley to grow, and today most of its surface area is of Holocene origin although it is evident that there was also a small BIA at LGM (Sinisalo *et al.* 2007).

Based on the $\delta^{18}O$ analysis (see section 4.2.1 for details), a high-elevation BIA in “Slöret” also originates in Holocene (Sinisalo 2007). “Slöret” is an unofficial name of a type III BIA located in the vicinity of the Slöret nunatak in Dronning Maud Land (see Fig. 1 for location). The results indicate that type III BIAs are also sensitive to the changes in the ice sheet thickness (Sinisalo 2007) although the mechanism of the formation of the BIA in “Slöret” is different from Scharffenbergbotnen. Delisle & Sievers (1991) found in their radio-echo sounding study in Allan Hills that BIAs can form if ice thickness is reduced to about 350 m or less. Ground penetrating radar (GPR) results show that the present day thickness of the "Slöret" BIA varies between 100 and 350 m (Sinisalo 2007). The ice flow over the nunataks must have slowed down as the ice sheet elevation decreased after LGM allowing a BIA to form. A similar condition is likely to be required for all open BIAs.

Interestingly, radar internal reflection analysis in central East Antarctica in the foreground of the Transantarctic Mountains (Fig. 1) suggests that ablation may have been more prevalent during glacial periods than in present (Siegert *et al.* 2003). Thus more and larger BIAs could have existed in the last glacial in this part of East Antarctica. This agrees with an overall Holocene increase in elevation of the East Antarctic ice sheet due to increased accumulation rates (Ritz *et al.* 2001), and highlights the differences between the ice sheet elevation history in the mountainous areas of Dronning Maud Land and in central East Antarctica and Transantarctic mountains.

4 Dating of blue ice

The principal problem in paleoclimate interpretation of blue ice samples has been dating the ice, as it is much more problematic than that of deep cores. None of the individual dating methods we shall discuss provide a reliable continuous dating over the surface of a BIA. They instead either date a specific layer e.g. dating of tephra deposits, or give an estimate of the age distribution of the overall surface as in most of the modelling efforts.

Dating a BIA can be resolved into two problems: 1) determining the horizontal age gradient along an ice flow line in the BIA – essentially the dip angle of the isochrones, and 2) finding the absolute age of ice at any particular reference point on the flow line. An absolute age marker may not be needed if it is possible to count annual layers from the mass balance equilibrium line. Its location, however, is variable on all time scales and it may be far up slope of the firn line marking the beginning of the BIA. Satisfactory dating requires both geophysical investigation and laboratory analyses before extraction of useful paleoclimate data from a BIA.

Opportunistic or reconnaissance age estimates of the BIAs have been largely based on the radiometric dating: terrestrial ages of meteorites collected from their surface of BIAs (Cassidy *et al.* 1992, Corti *et al.* 2003, Delisle 1993); ^{14}C in gas and microparticles within ice samples (Fireman 1986, van der Kemp *et al.* 2002, van Roijen *et al.* 1995, van Roijen 1996); and englacial tephra layers (Dunbar *et al.* 2008, McIntosh & Dunbar 2004, Wilch *et al.* 1999). Continuous dating requires much more extensive studies including ice flow modelling (e.g. Grinsted *et al.* 2003), stable isotopic values (Sinisalo *et al.* 2007), and stratigraphic comparisons with well-dated deep cores (Custer 2006, Moore *et al.* 2006). Geomorphology evidence of moraines provides a valuable addition to both reconnaissance and flow modelling dating methods. The best results are obtained by using a combination of methods as a single method alone cannot usually provide unambiguous dating.

4.1 Radiometric dating methods

4.1.1 Meteorites

The terrestrial ages of meteorites found on a BIA can be used as a measure of the age of the surface blue ice. The terrestrial ages of Antarctic meteorites determined using radioactive cosmogenic nuclides exceed two million years (Welten *et al.* 1995, Scherer *et al.* 1997), but are usually less than 500 thousand years (Welten *et al.* 2000).

Meteorites have been found on the surface of many Antarctic BIAs, most of which are located in the Transantarctic Mountains (Whillans & Cassidy 1983, Cassidy *et al.* 1992, Corti *et al.* 2003, Delisle 1993, Welten *et al.* 2000). No meteorites have been found in the mountainous BIAs in Dronning Maud Land. Possible explanations for this lack are:

1) young age of the BIAs (Sinisalo 2007),
2) meteorites may have melted back into the blue ice forming cryoconite holes (Fig. 4) during the high insolation days during summer, or
3) meteorites may be camouflaged by supraglacial debris on the surface of a BIA (Fig. 2, Hätterstrand & Johansen 2005).

Usually the meteorites have fallen on the snow accumulation areas that feed BIAs. These areas can be significantly larger than the actual BIAs, and hence they are more efficient collectors of samples than the BIAs. The meteorites are transported englacially to the BIA where they come to the surface along the ice flow, and they remain at the surface as ice around them ablates away, although sometimes winds can blow meteorites downstream along the ice surface. Thus, if the meteorite remains located where it came to the surface, while the ice continues to sublimate away, the ice is of same age or younger than the meteorites. If the meteorites are blown downstream then the ice may be older than the meteorite.

The accuracy of dating meteorites is about 30 kyr at its best (Welten *et al.* 2000). Goldstein *et al.* (2004) found meteorites of about 30 kyr terrestrial age in the same area at Allan Hills BIA where previously meteorites were found with ages more than 200 kyr. This shows that ice at the surface of the BIA can be significantly younger than the youngest meteorites dated in the area. It is also possible that meteorites land directly on the surface of a BIA as a result of direct infall (Huss 1990). Then the terrestrial age of the meteorite would be a measure of the minimum age of the BIA since the meteorite had never been buried. This is, however, not very likely as the snow covered accumulation areas of BIAs are usually much larger than the BIAs themselves.

The stochastic and labour-intensive nature of meteorite-collection, together with the non-unique relationship between their ages and that of the ice they are found on, means that they can only provide secondary dating rather than primary dating for a horizontal ice core transect. Progress must be also made in understanding the ice flow regime and long-term stability of the particular BIA to understand the ages better.

4.1.2 ^{14}C gas dating

Van Roijen (1996, van Roijen *et al.* 1994) developed a method for dating blue ice by measuring ^{14}C concentrations from air trapped in the ice, originally studied by Fireman and

Norris (1982). In van Roijen's method, ^{14}C depth profiles in blue ice are translated into carbon ages with a correction made for ^{14}C produced *in situ*, requiring an ice core to be analysed to about 50 m depth. The radiocarbon ages are converted to calendar ages using radiocarbon calibration curves (Reimer *et al.* 2004). The first BIA ^{14}C ages had large uncertainties of up to several thousands of years (van Roijen 1996). Van der Kemp *et al.* (2002) improved the method and decreased the uncertainty to ±400 years. This method, however, requires large ice samples of several kg and continuous ice core sections ideally at least 0.5 m long. This method does not provide an age profile with depth, simply a single age. The equipment and time required to drill ice cores on the BIA limits the sampling density achievable in a field campaign.

4.1.3 ^{14}C micro-particle dating

Recently Jenk *et al*. (2007) developed a technique to extract carbonaceous particles from ice core samples. In this method, dating requires only a few micrograms of carbon per sample, or about 1-5 kg of ice. Organic carbon is separated from elemental carbon and ^{14}C analysis is done by accelerated mass spectrometry. Ice cores from the Alps have been successfully dated (Jenk *et al*. 2006), with errors of about 2-10% over the last 500 years. The method shows a great promise for dating problematic traditional ice cores, and for BIAs. To date only one sample from a BIA has been studied, yielding a date for ice on Scharffenbergbotnen that was very consistent with other dating methods (see section 6.2).

4.1.4 Dating of tephra layers

Volcanic ash or tephra layers extracted from blue ice samples collected from Allan Hills (Goldstein *et al.* 2004, Fireman 1986), Yamato Mountains (Nishiizumi *et al.* 1979), and at Mount Moulton (Dunbar *et al.* 2008, Wilch *et al.* 1999, McIntosh & Dunbar 2004) (see Fig. 1 for the locations) have been radioisotopically dated by uranium-series and cosmogenic nuclides. In addition, grain size analysis of volcanic ash fragments can provide information about the distance to the source area (Dunbar *et al.* 2008, Nishio *et al.* 1984).

Boudinage may cause problems with radiodating of the material. The rheology of ice containing tephra (and soluble salts deposited with the tephra) is different from that of the surrounding cleaner ice. This induces flow instability that leads to pinching of layers of different "viscosity", even when the layers are under the pure shear stress. The wavelength of the boudins increases with decreasing viscosity contrast (Patterson 1994, p.182). For example, large piles of tephra are present on the surface of the ice at Mount Moulton located close to the Marie Byrd Land volcanoes. It makes it easy to sample collection for dating, but within a few tens of meters the layer that produced a large pile can thin to invisibility as a result of boudinage or snow patches especially in the older parts of a BIA where layers may be thinner. Therefore, it may be difficult to follow the tephra layers from a deposit large enough to date, to the place where a blue ice samples are taken.

Boudinage may also explain why the bands seen in other BIAs in Dronning Maud Land disappear or become hard to follow over longer scales, typically several hundred meters, as these bands contain much less material than Mount Moulton bands, and hence should have lower viscosity contrasts with clean ice. Large differences in results from uranium-series dating have been found for similar blue ice samples from Allan Hills (Goldstein *et al.* 2004, Fireman 1986). The reasons are not entirely clear but may be due to differences in sample processing (Goldstein *et al.* 2004).

4.2 Other reconnaissance methods

4.2.1 Isotopic dating

The ratios of heavy to light atoms of both oxygen and hydrogen in ice and snow, expressed as $\delta^{18}O$ and δD values, respectively, provide a simple method to determine whether samples at a given site were deposited during a glacial or an interglacial simply from their isotopic composition. The isotopic variability during the Holocene and glacial periods are known (e.g. Epica Community members 2006, Petit *et al.* 1999). A rapid, relatively large change of in δD or $\delta^{18}O$ in Antarctic ice is an indicator of a change between interglacial and glacial climates (e.g. Masson *et al.* 2000). Snow/firn samples from the accumulation area of a BIA represent the modern day isotopic values and isotopic values of the surface blue ice samples may be from glacial or interglacial periods. Possible changes in the ice sheet thickness in the last glacial-interglacial cycle must be taken into account especially in the low elevation mountainous areas where variations in the ice sheet thickness have been greatest (Pattyn & Decleir 1998). One complication in a simple interpretation of isotopic change as climate signature, is that the surface elevation of BIA is poorly constrained over time.

Results from the isotopic analysis of two BIAs in Dronning Maud Land are shown in Figure 5. Scharffenbergbotnen is a closed BIA (Grinsted *et al*. 2003), and it represents type I BIA in the classification of BIAs by Takahashi and others (1992) as mentioned earlier. “Slöret” is an open BIA of type III located at the edge of the Antarctic plateau at about 2300 m a.s.l. where the ice sheet flows over the Kirwanveggen mountain range. Based on the isotopic analysis (Fig. 5), it is obvious that the major part of the surface ice of these two mountainous BIAs are of Holocene origin. The natural limitation of this method is that it can only be used to make the division between samples of glacial and interglacial origin as described above.

4.2.2 Geomorphology

The supraglacial moraine structures can provide useful information about the age of a BIA. Hättestrand & Johansen (2005) studied moraines in Scharffenbergbotnen and Buråsbotnen BIAs in Dronning Maud Land (Fig. 6, see Fig. 1 for the location). They found that supraglacial moraines were deposited in Scharffenbergbotnen BIA during the LGM. They concluded that the survival of these moraines in the area indicates that there was a local ablation centre, and probably a BIA in the Scharffenbergbotnen valley at LGM, and it has been continuously maintained since then. The moraines on the slopes of surrounding nunataks suggest that the surface elevation of the BIA has been higher during LGM, and striae and gouges in the outcropping bedrock indicate past ice flow directions (Hättestrand & Johansen 2005). The moraines drawn out from the Buråsbotnen cirque (Fig. 6), on the other hand, have been deposited in the inner part of the cirque in an earlier phase, but subsequent increased accumulation in the area has caused them to be pushed and drawn out in the overall ice flow direction towards the north-west (Hättestrand & Johansen 2005). Similarly, old supraglacial moraines and ice flow indicators in the bedrock in other BIAs can be used for reconstructing the former ice flow, and existence, surface elevation, and extent of the BIA.

4.3 More intensive methods

4.3.1 Ice flow modelling

In contrast with the methods outlined so far in sections 4.1 and 4.2, ice flow modelling produces a continuous set of ages over a BIA, or at least along selected flow lines on a BIA. A full three dimensional model is always the best approach for ice sheet modelling. The necessary input field data, however, are generally unavailable, or derived from short period observations. Typical ice flow velocities are 10 cm a^{-1} in both horizontal and vertical components, and this is lower than on many ice sheets, hence it is often challenging to collect the appropriate data sets on BIAs. Thus, it is much more practical to use a flow line approach than a three dimensional model, as sparse field data can be more easily incorporated into such a model.

Traditional flow line models have either been based on the shallow ice approximation (e.g. Paterson 1994) or the full stress equilibrium equations (e.g. Pattyn 2002). The shallow ice type models have problems in BIAs where the flow is not always down slope, bedrock relief is large or the flow is constrained by valley sides. Models based on the full stress equilibrium equations are very good for predicting how the ice reacts to mass balance perturbations. However, these models are not fully constrained to the observed data.

The first modelling efforts on BIAs were made to explain the meteorite findings on the surface of BIAs. Nasure & Hashimoto (1982) made a simple flow model to date a BIA upstream of a nunatak in Yamato Mountains based on the continuity equation (Nye 1953). Whillans & Cassidy (1983) produced a very simple model for the Allan Hills BIA, and a much more sophisticated approach was used by Azuma *et al.* (1985) for the flow on South Yamato BIA in an area near a nunatak called Massif A. Azuma *et al.* (1985) used a 30 m ice core taken 2 km upstream of the nunatak to estimate parameters for the flow law of the ice, and then computed flow lines for the ice near the nunatak, parameterizing the flow divergence near the nunatak. In common with these earlier modelling efforts, van Roijen (1996) parameterized ice thickness but used some limited velocity observations and ^{14}C ages to constrain flow.

Grinsted *et al.* (2003) developed a volume conserving flow line model which assumes constant ice sheet geometry over time, i.e. steady state flow, and which avoids the need to parameterize flow divergence. This approach allows use of the full set of relatively sparse observational data to simulate realistic shearing flow of the ice sheet, where we have no measurements of the temperature profile, and hence velocity variation with depth. This model simply uses volume conservation forced by observed surface velocities, ice thickness and mass balance. Hence there is no time evolution of the glacier surface elevation, and all force imbalance is taken up in the vertical velocity which is a model output. A useful feature of the model is that the vertical velocity profile is parameterized in terms of a temperature dependent ice rheology that depends primarily on the shape of the temperature profile scaled by the surface velocity. The steady state temperature profile in a cold-based ice sheet is linear at the equilibrium line between net accumulation and ablation, and as we are dealing with slow flow on each side of the equilibrium line a linear profile may be reasonable for the whole flow line in BIAs. This model has been applied to Scharffenbergbotnen (Grinsted *et al.* 2003, Sinisalo *et* *al.* 2004, 2007), Allan Hills (Grinsted *et al.* 2003) and to South Yamato BIAs (Moore *et al.* 2006).

The great disadvantage of the modelling to date is that it is essentially prognostic – that is time-independent, and the Grinsted *et al*. (2003) model being volume conserving precludes any ice sheet elevation change. Over glacial-interglacial cycles ice sheet elevation changes must have played a significant role in ice flow at most BIAs. Additionally, it is still unclear how much of the bed of BIAs are frozen. The surface velocities are not necessarily directly related to creep flow alone. The mountain BIAs are situated in the area of the cold/warm transition at the bed as the ice sheet is mainly cold based on the Antarctic plateau, while the ice is at the pressure melting point downstream of the mountains bordering the East Antarctic ice sheet (Huybrechts 1992). The melting condition at the bed is essential for modeling the evolution of BIAs.

We have experimented with a three-dimensional fully coupled thermomechanical finite element model (ELMER-ICE) of Scharffenbergbotnen. Input to these models requires the geometry of glacier upper and lower boundaries and the mass and energy and fluxes across each boundary. These data are now reasonably well known for Scharffenbergbotnen, but to our knowledge is lacking for other BIAs. Output from such models includes basal temperatures (with the proviso that the geothermal heat flux must be input, or at least constrained by very sparse available data). Results show encouraging prognostic behaviour that matches the well measured ice velocity patterns. Development of the diagnostic, time evolving model is in progress.

4.3.2 Radar isochrones

BIA geometry is a vital input to modelling, and that is most easily found by radar sounding the BIA. These radar data can also include internal layering of the snow/blue ice transition zone (Sinisalo *et al.* 2004), which can give valuable information on accumulation rates and isochrone dip angles can be used to validate flow model predictions. In some BIAs the dip angle has been measured directly. For example, the observed dip angle of the blue ice relative to the surrounding firn is only a few degrees on South Yamato BIA (Yokoyama 1976) while the dipping angles of tephra bands in Mount Moulton are typically about 45° (Dunbar *et al.* 2008). A relatively low dip angle is an inevitable consequence of open type BIAs while the higher angles are found in closed BIAs.

Sinisalo *et al*. (2004) showed at the dip angle of an isochrone can be calculated from the along-flow surface velocity profile *u* and the ablation and accumulation gradients *b'* and *a'* by considering the layer geometry of the along-flow surface-age ($d\tau/dx$) and vertical age ($d\tau/dz$) gradients. Moore *et al*. (2006) showed that this angle was also simply related to the accumulation rate *a* along the flow line, if the amount of strain thinning along the flow line can be estimated such that *a* at the source region is thinned to λ due to flow. These then yield relationships for the dip angle ϕ

$$\tan(\phi) = \frac{\left(\dfrac{\partial \tau}{\partial x}\right)}{\left(\dfrac{\partial \tau}{\partial z}\right)} = \frac{-x}{u}\left(\sqrt{a'b'} + b'\right) = \frac{\lambda - b}{u} \qquad (1)$$

The two right hand side equalities of Equation (1) are useful in different circumstances. The equality showing the linear dependence of the dip angle on the horizontal distance *x* from the equilibrium line is useful near the equilibrium line. That is where radar data often show isochrones well, and where strain thinning of the layers can be neglected. The dependence on *x* predicted from (1) is consistent with the GPR data that show near surface layers less steep than the deeper ones (Fig. 7). The strain thinning introduced in the rightmost equality is more useful far from the equilibrium line where dip angles are occasionally available, such as tephra layers. Equation (1) assumes that the ablation rate is constant over time. However, these functions need not to be constant over time and detailed analysis of radar profiles such as in Figure 7 together with flow modelling should enable details of the ice accumulation pattern in the region to be well constrained.

Equation (1) shows that for an open BIA (e.g. South Yamato) flow, where *u* remains relatively high across the BIA, the dip angle will be low. One relevant implication of shallow dip angles is that although the ice is ancient at the surface, a horizontal ice core is not a particularly efficient way of recovering ice spanning a great age. This can also be appreciated from the roughly 50 km along flow extent of the BIA that spans about 150 kyr compared with 150 kyr spanned by the upper 1.8 km of the Dome Fuji ice core (Watanabe *et al.* 2003b).

4.3.3 Stratigraphic matching

A traditional method of dating stratigraphic records such as paleoclimate proxies is by cross-matching them an established well-dated record. To do this requires that the proxies being compared are responding to the same climate variability, and not responding to some local variability that is unrepresented in the reference profile. The most useful stratigraphic paleoclimate records appear to be the relative concentration and isotopic composition of the gases trapped in the ice. As gases are rapidly mixed in the atmosphere even records from Greenland can be compared with Antarctica. However, there are problems in collecting samples suitable for gas analysis from BIAs, most likely due exchange of gases in the upper metres of the blue ice which experiences cracking as it approaches the surface. The work on bubble expansion from a 100 m core from South Yamato BIA (Nakawo *et al.*1988) suggests that significant release of stress occurs even at 50 m depth, though no cracks were seen below 7 m depth. The relatively small upward velocities in BIAs of order 10 cm a^{-1} mean that ice is in the upper 10 m of the BIA for century periods of time.

Stratigraphic matching may also be unreliable due to compositional changes as the ice sits in a near surface environment. The rate at which the ice approaches the surface is comparable with the firnification rates on the Antarctic plateau, where post-depositional changes can occur (Legrand & Mayewski 1997). This also leads to the problems in ^{14}C dating of near surface air bubbles ice caused by in-situ ^{14}C production by cosmic-ray muons (van der Kemp *et al.* 2002). The presence of cryoconite and biological activity in the ice can also lead to chemical conversion of species within the upper decimetre or so of ice. This can be avoided by sampling deep enough, but that leads to practical difficulties. The chemical species that appears to be most affected by near-surface processes is MSA. In profiles of 10 m cores in BIA we find no clear sign of any trend with any other species.

Volcanic signatures are potentially a useful method of cross-matching records extracted from BIAs and deep ice cores. However, the individual analysis of tephra by electron microscopy methods is a demanding and laborious method (e.g. Kekonen *et al*. 2005, Palais *et al.* 1992), and it is not suitable for continuous profiling at present. Proxy methods of identifying volcanic signatures from sulphate profiles (Zielinski 1995), or from their electrical conductivity signature (Hammer *et al.* 1980) are problematic as the volcanic fallout is never uniform across the ice sheet, and there are significant volcanic acid fallout events rather frequently (Robock & Free 1995). This leads to difficulties in finding a unique match of a series of spikes in a BIA record with a reference ice core (Moore *et al.* 2006).

The process of wiggle-matching using ice parameters such as stable water isotope, or major soluble ions between a BIA record and a deep ice core is perhaps more reliable, though potentially undermines the point of extracting the BIA record. The advantage of the BIA record is that it is from a different geographic area than the deep ice cores, hence we would expect to find differences in the climatic signatures between the two records. Forcing the main variability to match between deep cores sites and BIAs reduces the chances of seeing genuine climatic variability recorded in the two records.

4.3.4 Layer counting

A plausible direct method of dating is by counting cycles in measured parameters. This is potentially highly useful on BIAs where the annual layer thickness is greater than found deep in traditional ice cores, both because annual accumulation rates nearer the coast are higher than inland, and because the amount of strain thinning experienced is often less. Therefore the number of samples per cycle is much less limited by sample size than in vertical deep cores. The annual layer thickness at mid-Holocene measured from a single 60-cm sample at the surface of the Scharffenbergbotnen BIA, for example, is about 20 cm of ice that corresponds to 160 mm w.e., and to a horizontal age gradient of about 5.4 a m^{-1} (Fig. 8, Sinisalo *et al.* 2007).

BIA records from coastal areas may have less well-defined annual cycles than seen in more inland records, however, especially in those from largely sea-salt derived species (Legrand & Mayewski 1997). As with cycle counting in vertical cores, the greater the number of different annually-varying parameters that are considered, the more accurate will be the annual cycle count (Taylor *et al*. 2004).

Specific difficulties with BIA layer counting are firstly the dip angle of layers. If the dip angle is too shallow it is quite possible that the small-scale topography of the ice surface will re-sample the same layer more than once along the transect. The second problem is the need for an absolute date to anchor the cycle count. Thus, it is possible that a rather accurate number of cycles could be found in a given part of the BIA, but actual age of the ice remains only known to the limits of radiometric or ice flow precision. There are practical difficulties in determining the equilibrium line, and hence zero age of the BIA. The equilibrium line will be some distance up flow from the firn line, and the upstream part of the surface ice zone is marked with old snow and firn patches that make surface sampling virtually impossible. One possible method for some BIA would be high frequency radar to count layers in the equilibrium area and firn areas until continuous surface sampling of ice could begin.

4.4 Blue ice cores

Very few blue ice cores have been studied for paleoclimatic purposes to date (Table 1). There are only two horizontal ice cores analysed for paleoclimate records: a 101-m long ice transect from the surface of the Scharffenbergbotnen BIA (Sinisalo *et al.* 2007), and a 600-m long transect from Mount Moulton (Popp *et al.* 2004). The transect in Scharffenbergbotnen was estimated to represent a 500-year period in mid-Holocene whereas the Mount Moulton does not include the Holocene period since that part of the BIA was covered by snow when sampling was done.

The temporal resolution of the paleoclimate records extracted from blue ice cores vary greatly. Isotopic analysis of a sample from the horizontal transect in Scharffenbergbotnen suggested a horizontal age gradient of about 5 a m^{-1} corresponding to about 160 mm w.e. (Fig. 8, Sinisalo *et al.* 2007). The gradient is fairly linear along the flowline over the BIA, and it covers the entire Holocene period. The stable isotope record from the transect in Mount Moulton, on the other hand, spans about 150 000 years within 600 m. It is the first such record of the last interglacial in West Antarctica, and it shares many features with the records from deep ice cores from Vostok and EPICA-Dome C in East Antarctica (Popp *et al.* 2004). Popp *et al.* (2004) found also that the trapped gases in the blue ice samples correlated well with those of Vostok ice core.

In addition to the horizontal transects, vertical blue ice cores of 30-101 m have been drilled for paleoclimate studies in Mount Moulton, Scharffenbergbotnen and Yamato Mountains. The age of the upper part of the vertical blue ice core drilled in Scharffenbergbotnen is about 11 kyr defined by ^{14}C gas dating (van der Kemp *et al.* 2002). The 30-m blue ice core from Mount Moulton includes a gas record between the ages of ~117-135 kyr (Custer, 2006) whereas the 101 m vertical blue ice core from South Yamato BIA contains a section spanning about 6 kyr from 60 kyr BP (Moore *et al.* 2006). The Yamato Mountain blue ice core was dated by taking advantage of the great wealth of previously unpublished and published data on South Yamato BIA that were used as inputs and constraints in a flow line model (Moore *et al.* 2006). The geophysical data (Lythe *et al.* 2001, Takahashi *et al.* 1994) allowed construction of a flow line from the ice core on the BIA to the deep borehole at Dome Fuji. In order to determine the age span, the gas data measured from the 101-m blue ice core (Machida *et al*. 1996) were fitted with records from Dome Fuji and other ice cores in both polar regions (Watanabe *et al*. 2003a, Kawamura *et al*. 2003, Blunier & Brook 2001, Sowers *et al*. 2003). More precise dating was achieved with high-resolution records of electrical stratigraphy from Dome Fuji ice core. This approach of combination of methods can be used quite generally to link deep ice cores to surface outcrops on blue ice fields.

4.5 Recommendation for blue ice dating

Greatest confidence in dating blue ice comes by combining many dating methods together. Generally, this means tuning flow models with other methods. First, a flow model can be used to gain an initial expectation of a plausible surface age distribution, vertical age span, and the source region of the ice. Other dating methods, such as radiocarbon dating, meteorite terrestrial ages, or tephra layer dating can be used with the model in an iterative way to estimate the age distribution in a BIA. The flow modeling combined with other methods (Fig. 10) can also provide surface age distribution along the flow line, temporal changes in the size of the BIA, and ice flow velocities, and constrain elevation changes and accumulation patterns over the accumulation region.

5 General remarks on blue ice areas

It is clear from the investigations to date that BIAs in different parts of Antarctica have very different histories. The area covered by blue ice, the number of individual BIAs, and their geographical distribution vary mainly on an ice sheet thinning-thickening timescale. Many BIAs in Dronning Maud Land are of Holocene age whereas the South Yamato BIA seems to be rather stable, having been only about 10% smaller in the glacial than at present (Moore *et al.* 2006). Large parts of the central East Antarctica, on the other hand, may have been ablation zones during the last glacial (Siegert *et al.* 2003), consistent with general lowing of accumulation and higher wind speeds during colder climate periods. Isolated BIAs, such as Mount Moulton sitting at high elevation on a volcanic peak where the BIA is created by a nunatak only about 10 m higher than present ice cap, are likely to respond to more localized climate than the BIAs in the mountains fringing East Antarctica. In general, the BIAs accompanying the highest and steepest nunataks are most likely to be stable (Bintanja 1999).

The BIAs show often very different physical characteristics, and temporal resolution of the paleoclimate records extracted from blue ice cores vary greatly. Figure 11 illustrates the variability in the surface age distributions, and lengths of some BIAs along their flow lines. The large South Yamato field has long flow lines associated with it (typically hundreds of kilometres), while Allan Hills and many in Dronning Maud Land collect ice from only a few kilometres distance. The measured surface velocities vary from the few centimeters (e.g. Dunbar *et al.* 2008) per year to several metres per year (e.g. Moore *et al.* 2006). As a result, the surface age gradients of BIAs can be of different magnitudes.

6 Summary and suggestions for future studies

Blue ice areas show potential for paleoclimate research applications while very little paleoclimate data from BIAs have been published to date. Blue ice samples collected from BIA surface may provide paleoclimate data with higher resolution than traditional deep ice cores. In addition, many BIAs are located in areas where no deep ice cores are available. Dating of BIAs remains still challenging, but as the examples in this paper show, substantial progress has been made in that field. Blue ice cores and ice flow modelling combined with other dating methods have provided insight in the variability of dynamics and sizes of the BIAs, and the surrounding ice sheet, and local climatic conditions. The conditions that create and maintain BIAs are of wide significance as they are related to the general ice sheet elevation, accumulation patterns and climate.

BIAs located in different sectors close to Southern Ocean can provide high resolution climatic information about the last glacial transition, as the Southern Ocean may trigger the transition into an interglacial mode of circulation (Knorr & Lohmann 2003). Many BIAs are located much closer to the ocean than traditional deep coring sites are, and they are relatively easy to access.

The mountain BIAs are situated in the area of the cold/warm transition at the bed. The East Antarctic ice sheet is mainly cold based on the plateau, while ice is at the pressure melting point downstream of the mountains bordering the ice sheet (Huybrechts 1992). The bed conditions of BIAs should be studied more intensively as it is essential for modeling the evolution of BIAs, and also for detailed modeling of the larger ice sheet.

Satellite and aerial photographic image analyses of BIAs can provide new information about the dynamics of the Antarctic ice sheet. Correlation of individual, dated tephra layers, or sets of layers, in BIAs will allow a better understanding of the geometry of the ice flow. The paleoclimate record obtained from surface blue ice samples can be extended over wide geographical areas, though at low resolution, by using satellite images. This can be made by identifying and dating visible horizons on BIAs using the stratigraphic records from ice cores.

Satellite imagery shows potential to be a valuable tool in monitoring the possible areal changes in individual BIAs, and in the total extent of Antarctic BIAs. Although the overall changes in the BIA extent seem to be slow, BIAs should be continued to be monitored by satellite imagery, especially using radar images which may allow near surface blue ice to be observed under *in-situ* snow cover.

7 Acknowledgements

We would like to thank A. Grinsted for comments that helped us to improve the manuscript. Finnish Antarctic Research Program (FINNARP 1999-2000, 2000-2001, and 2003-2004) provided the logistics for the field work. In addition, many Antarctic teams provided logistical support and data collection, especially, SANAE 39 and 40, and JARE 24 and 27. This work was primarily funded by the Academy of Finland and the Thule Institute. Parts of this work were funded by the Arctic graduate school ARKTIS, and by grants from the Faculty of Natural Sciences, Oulu University and the University Pharmacy Foundation (Oulu).

8 References

Azuma, N., M. Nakawo, A. Higashi, and F. Nishio, 1985. Flow pattern near Massif A in the Yamato bare ice field estimated from the structures and the mechanical properties of a shallow ice core, *Mem. NIPR, Spec. Iss. 39*, 173-183.

Bintanja, R., 1999. On the glaciological, meteorological, and climatological significance of Antarctic blue ice areas. Rew. Geohys., 37,3, p. 337-359. (1999RG900007).

Bintanja, R., and C. H. Reijmer, 2001. Meteorological conditions over Antarctic blue-ice areas and their influence on the local surface mass balance. *J. Glaciol.,* 47 (156), 37-50.

Bintanja, R., C. H. Reijmer, and S. J. M. H. Hulscher, 2001. Detailed observations of the rippled surface of Antarctic blue-ice areas. *J. Glaciol*., Vol. 47, No. 158, 387-396.

Bintanja, R., and M. R. van den Broeke, 1995a. The climate sensitivity of Antarctic blue ice areas, *Ann. Glaciol.*, 21, 157-191.

Bintanja, R. and M. R. van den Broeke, 1995b. The surface energy balance of Antarctic snow and blue ice. *J. Appl. Meteor.*, 34, 902-926.

Blunier T., and E. J. Brook, 2001. Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science,* 291, 109-112.

Brown, I. C., and T. A. Scambos, 2004. Satellite monitoring of blue-ice extent near Byrd Glacier, Antarctica. *Ann. Glaciol.* 39, p. 223-230.

Cassidy, W., R. Harvey, J. Schutt, G. Delisle, and K. Yanai, 1992. The meteorite collection sites of Antarctica. *Meteoritics* 27, 490–525.

Cassidy, W. A., E. Olsen, and K. Yanai, 1977. Antarctica: A Deep-Freeze Storehouse for Meteorites. *Science* 198 (4318), 727. [DOI: 10.1126/science.198.4318.727].

Coren, F., G. Delisle, and P. Sterzai, 2003. Ice dynamics of the Allan Hills meteorite concentration sites revealed by satellite aperture radar interferometry. *Meteoritics and Planetary Science* 38 Nr 9 (2003) 1319-1330.

Corti, G., A. Zeoli, P. Belmaggio, and L. Folco, 2008. Physical modeling of the influence of bedrock topography and ablation on ice flow and meteorite concentration in Antarctica. *J. Geophys. Res.*, 113, F01018, doi:10.1029/2006JF000708.

Corti, G., A. Zeoli, and M. Bonini, 2003. Ice-flow dynamics and meteorite collection in Antarctica. *Earth and Plan. Sci. Lett*. 215 (2003) 371-378.

Crary, A. P. and C. R. Wilson, 1961. Formation of “Blue” Glacier Ice by Horizontal Compressive Forces. *J. Glaciol.*, Vol. 3, No. 30, 1045-1050.

Custer, S. E., 2006. Eemian records of ∂^{18}Oatm and CH_4 correlated to the Vostok EGT4 timescale from the Moulton Blue Ice Field, West Antarctica. *A senior thesis in Geosciences*, The Pennsylvania State University, USA. http://www.geosc.psu.edu/undergrads/documents/documents/StantonCusterthesis.pdf.

Delisle, G. 1993. Global change, Antarctic meteorite traps and the East Antarctic ice sheet. *J. Glaciol*., 39, 397–408.

Delisle, G. and J. Sievers. 1991. Sub-ice topography and meteorite finds near the Allan Hills and the Near Western Icefield, Victoria Land, Antarctica. *J. Geophys. Res*., E96, 15577-15587.

Dunbar, N. W., W. C. McIntosh, and R. P. Esser, 2008. Physical setting and tephrochonology of the summit caldera ice record at Mt. Moulton, West Antarctica. Geological Society of America Bulletin, v. 120, no. 7/8, p. 796-812. doi:10.1130/B26140.1.

EPICA Community Members, 2006. One-to-one hemispheric coupling of millennial polar climate variability during the last glacial. *Nature*, 444, 195-198.

Faure G., and D. Buchanan, 1991. Ablation rates of the ice fields in the vicinity of the Allan Hills, Victoria Land, Antarctica. Contributions to Antarctic research II. *Antarctic Res. Ser.*, Vol 53, pp. 19-31.

Fireman, E. L., 1986. Uranium-series dating of Allan Hills ice. *J. Geophys. Res*. 91, pp. D539–D544 (correction J. Geophys. Res. 91: 8393).

Fireman E. L. and T. L. Norris, 1982. Ages and composition of gas trapped in Allan Hills and Byrd core ice. *Earth and Planetary Science Letters*, Volume 60, Issue 3, 339-350. doi:10.1016/0012-821X(82)90072-3.

Giaever, J., 1954 *The White Desert*. The Official Account of the Norwegian- British-Swedish Antarctic Expedition. Chatto & Windus, London 1954.

Goldstein S. J., M. T. Murrell, K. Nishiizumi, and A. J. Nunn, 2004. Uranium-series chronology and cosmogenic 10Be-36Cl record of Antarctic ice. *Chemical Geology*, 204, 125-143.

Grinsted, A., J. C. Moore, V. Spikes, and A. Sinisalo, 2003. Dating Antarctic Blue Ice Areas using a novel ice flow model. *Geoph. Res. Lett*. 30(19), 2005. (10.1029/2003GL017957).

Hammer, C. U., H. B. Clausen, and W. Dansgaard, 1980. Greenland ice sheet evidence of post-glacial volcanism and its climate impact. *Nature*, 288:230.

Harvey, R. P., 2003. The origin and significance of Antarctic meteorites. *Chemie der Erde, 63, 93-147*. DOI: 10.1078/0009-2819-00031.

Hättestrand, C., and N. Johansen, 2005. Supraglacial moraines in Scharffenbergbotnen, Heimefrontfjella, Dronning Maud Land, Antarctica –significance for reconstructing former blue ice areas. *Antarctic Science* 17 (2), 225-236 (2005). DOI: 10.1017/S0954102005002634.

Hodson, A. J., P. N. Mumford, J. Kohler, and P. M. Wynn, 2005. The High Arctic glacial ecosystem: new insights from nutrient budgets. *Biogeochem.*,**72**, 233 – 256.

Huss, G. R., 1990. Meteorite infall as a function of mass: Implications for the accumulation of meteorites on Antarctic ice. *Meteoritics* 25 (1990), 41-56.

Huybrechts, P., 1992. The Antarctic ice sheet and environmental change: a three-dimensional modeling study. *Ber. Polarforsch.* **99**, p. 1–241.

Jenk, T. M., S. Szidat, M. Schwikowski, H. W. Gaeggeler, S. Brütsch, L.Wacker, H.-A. Synal, and M. Saurer, 2006. Radiocarbon analysis in an Alpine ice core: record of anthropogenic and biogenic contributions to carbonaceous aerosols in the past (1650–1940). *Atmos. Chem. Phys.*, 6, 5381–5390.

Jenk T. M., S. Szidat, M. Schwikowski, H. W. Gaeggeler, L. Wacker, H.-A. Synal, and M. Saurer, 2007. Microgram level radiocarbon (^{14}C) determination on carbonaceous particles in ice. *Nuclear Instruments and Methods in Physics Research* B 259 (2007) 518–525.

Jouzel, J., G. Raisbeck, J.P. Benoist, F. Yiou, C. Lorius, D. Raynaud, J. R. Petit, N. I. Barkov, Y. S. Korotkevitch, and V. M. Kotlyakov, 1989. A comparison of deep Antarctic ice cores and their implications for climate between 65,000 and 15,000 years ago. *Quat. Res., 31,* 135-150.

Kawamura, K., T. Nakazawa, S. Aoki, S. Sugawara, Y. Fujii, and O. Watanabe, 2003. Atmospheric CO_2 variations over the last three glacial-interglacial climatic cycles deduced from the Dome Fuji deep ice core, Antarctica using a wet extraction technique. *Tellus B*, 55 (2), 126-137.

Kekonen, T., J. C. Moore, P. Perämäki and T. Martma, 2005. The Icelandic Laki volcanic tephra layer in the Lomonosovfonna ice core, Svalbard. *Polar Research* 24(1-2), 33-40.

Knorr, G. and G. Lohmann, 2003. Southern Ocean Origin for Resumption of Atlantic Thermohaline Circulation during Deglaciation. *Nature*, 424, 532-536

Koeberl, C., 1990. Dust Bands in Blue Ice Fields in Antarctica and Their Relationship to Meteorites and Ice. Workshop on Antarctic Meteorite Stranding Surfaces. A Lunar and Planetary Institute Workshop held 13-15, 1988, at the University of Pittsburgh. Sponsored by Division of Polar Programs, National Science Foundation, and LPI. Edited by W. A. Cassidy and I. M. Whillans. *LPI Technical Report 90-03*, published by Lunar and Planetary Institute, 3303 NASA Road 1, Houston, TX 77058, 1990, p.70

Koeberl, C., K. Yanai, W. A. Cassidy, and J. W. Schutt, 1988. Investigation of dust bands from blue ice fields in the Lewis Cliff (Beardmore) area, Antarctica: A progress report. *Proc. NIPR Symp. Ant. Meteor.* 1, 291-309.

Legrand, M. and P. Mayewski. 1997. Glaciochemistry of polar icecores: a review. *Rev. Geophys.*, **35**(3), 219–243.

Lythe, M. B., D. G. Vaughan and the BEDMAP Consortium, 2001. BEDMAP: A new ice thickness and subglacial topographic model of Antarctica. *J. Geophys. Res. 106,* 11335-11351.

Machida, T., T. Nakazawa, H. Narita, Y. Fujii, S. Aoki and O. Watanabe , 1996. Variations and the CO_2, CH_4 and N_2O concentrations and $\delta^{13}C$ of CO_2 in the glacial period deduced from an Antarctic ice core, South Yamato. *Proc. NIPR Symp. Polar Meteorol. Glaciol., 10,* 55-65.

Masson V., F. Vimeux, J. Jouzel, V. I. Morgan, M. Delmotte, P. Ciais, C. U. Hammer, S. J. Johnsen, V. Y. Lipenkov, E. M. Thompson, J-R. Petit, E. J. Steig, M. Stievenard, and R. Vaikmae, 2000. Holocene Climate Variability in Antarctica Based on 11 Ice-Core Isotopic Records. *Quat. Res.*, Vol. 54, 348 - 358.

McIntosh, W. C., and N. W. Dunbar, 2004, High-precision 40Ar/39Ar dating of a 10 ka to 492 ka sequence of englacial tephra layers at Mt. Moulton, Antarctica: IAVCEI International Volcanological Congress.

Mellor, M. and C. Swithinbank, 1989. Airfields on Antarctic glacier ice. *CRRELL Rep*. 89-21.

Moore, J. C., F. Nishio, S. Fujita, H. Narita, E. Pasteur, A. Grinsted, A. Sinisalo, and N. Maeno, 2006. Interpreting ancient ice in a shallow ice core from the South Yamato (Antarctica) blue ice area using flow modeling and compositional matching to deep ice cores. *J. Geophys. Res.*, *111*, D16302, doi:10.1029/2005JD006343.

Nakawo, M, M. Nagoshi, and S. Mae, 1988. Stratigraphic record of an ice core from the Yamato meteorite ice field, Antarctica, *Ann. Glaciol.*, *10*, 126-129.

Nakamura, T. N. Imae and I. Nakai *et al.*, 1999. Antarctic micrometeorites collected at the Dome Fuji Station, *Antarct. Meteorit. Res.* **12**, pp. 183–198.

Naruse, R., and M. Hashimoto, 1982. Internal flow lines in the ice sheet upstream of the Yamato Mountains, East Antarctica. *Mem. NIPR spec. Iss. 24*, 201-203.

Näslund, J. O., J. L. Fastook, and P. Holmlund, 2000. Numerical modeling of the ice sheet in western Dronning Maud Land, East Antarctica: impacts of present, past and future climates. *J. Glaciol.* 46 (152), 54-66.

Nishiizumi, K., D. Elmore, and P. W. Kubik, 1989. Update on terrestrial ages of Antarctic meteorites, *Earth Planet. Sci. Lett.*, 93, 299-313.

Nishiizumi, K., L. R. Arnold, D. Elmore, R. D. Ferraro, H. E. Gove, R. C. Finkel, R. P. Beukens, K. H. Chang, and L. R. Kilius, 1979. Measurements of 36CL in Antarctic meteorites and Antarctic ice using a Van de Graaff accelerator. *Earth. Planet. Sci. Lett.*, 45, 285-292.

Nishio, F., T. Katsushima, and H. Ohmae, 1985. Volcanic ash layers in bare ice areas near the Yamato Mountains, Dronning Maud Land and the Allan Hills, Victoria Land, Antarctica. *Ann. Glaciol.*, 7, 34-41.

Nishio, F., T. Katsushima, H. Ohmae, M. Ishikawa, and S. Takahashi, 1984. Dirt layers and atmospheric transportation of volcanic glass in the bare ice areas near the Yamato Mountains in Queen Maud Land and the Allan Hills in Victoria Land, Antarctica. *Mem. NIPR spec. Iss. 34*, 160-173

Nye, J. F., 1963. Correction factor for accumulation measured by the thickness of the annual layers in an ice sheet. *J. Glaciol.*, **4**(36), 785-788.

Orheim O., and B. Lucchitta, 1990. Investigating climate change by digital analysis of blue ice extent on satellite images of Antarctica. *Ann. Glac.* 14., p.211-215.

Palais, J.M., M. S. Germani, and G. A. Zielinski, 1992. Inter-hemispheric transport of volcanic ash from a 1259 A.D. volcanic eruption to the Greenland and Antarctic ice sheets. *Geophysical Research Letters 19*, 801-804.

Paterson, W. S. B., 1994. *The Physics of Glaciers*, 3rd ed. Pergamon.

Pattyn, F., 1999. The variability of Antarctic ice-sheet response to the climatic signal. *Ann. Glaciol.* 29, 273-278.

Pattyn, F. and H. Decleir, 1998. Ice dynamics near Antarctic marginal mountain ranges: implications for interpreting the glacial-geological evidence. *Ann. Glaciol.*, 27, 327-332.

Pattyn, F., 2002. Transient glacier response with a higher-order numerical ice-flow model. *J. Glaciol.*, 48, 467-477.

Perchiazzi, N., L. Folco, and M. Mellini, 1999. Volcanic ash bands in the Frontier Mountain and Lichen Hills blue-ice fields, northern Victoria Land. *Antarctic Science* 11 (3): 353-361.

Petit, J.R., J. Jouzel, D. Raynaud, N.I. Barkov, J.-M. Barnola, I. Basile, M. Bender, J. Chappellaz, M. Davis, G. Delayque, M. Delmotte, V.M. Kotlyakov, M. Legrand, V.Y. Lipenkov, C. Lorius, L. Pépin, C. Ritz, E. Saltzman, and M. Stievenard, 1999. Climate and Atmospheric History of the past 420,000 years from the Vostok Ice Core, Antarctica. *Nature* 399: 429-436.

Popp, T., T. Sowers, N. Dunbar, W. McIntosh, and J. W. C. White, 2004. Radioisotopically dated climate record spanning the last interglacial in ice from Mount Moulton, West Antarctica, *Poster* Presented at the AGU Fall Meeting, December 2004. American Geophysical Union, San Francisco.

Reimer, P. J., M. G. L. Baillie, E. Bard, A. Bayliss, J. W. Beck, C. J. H. Bertrand, P. G. Blackwell, C. E. Buck, G. S. Burr, K. B. Cutler, P. E. Damon, R. L. Edwards, R. G. Fairbanks, M. Friedrich, T. P. Guilderson, A. G. Hogg, K. A. Hughen, B. Kromer, G. McCormac,, S. Manning, C. Bronk Ramsey, R. W. Reimer, S. Remmele, J. R. Southon, M. Stuiver, S. Talamo, F. W. Taylor, J. van der Plicht, and C. E. Weyhenmeyer, 2004. INTCAL04 Terrestrial radiocarbon age calibration 0-26 cal kyr BP. *Radiocarbon* 46, 1029-1058.

Ritz, C., V. Rommelaere, and C. Dumas, 2001. Modeling the evolution of Antarctic ice sheet over the last 420,000 years: Implications for altitude changes in the Vostok region. *J. Geophys. Res.*, *106*(D23), 31,943–31,964.

Robock, A. and M.P. Free, 1995. Ice cores as an index of global volcanism from 1850 to the present. *J. Geophys. Res*., 100, 11,549-11,567.

Scherer, P., L. Schultz, U. Neupert, M. Knauer, S. Neumann, I. Leya, R. Michel, J. Mokos, M. E. Lipschutz, K. Metzler, M. Suter, and P. W. Kubik, 1997. Allan Hills 88019: an Antarctic H-chondrite with a very long terrestrial age. *Meteoritics & Planetary Science*, vol. 32, no. 6, pages 769-773

Schultz, L., J. O. Annexstad, and G. Delisle, 1990. Ice movement and mass balance at the Allan Hills Icefield. *Antarctic J., US* 25, pp. 94-95.

Schytt, V., 1961. Glaciology II. E. Blue ice fields, moraine features and glacier fluctuations. *Norwegian-British-Swedish Antarctic Expedition, 1949-52, Scientific results,* vol. IV E, pp. 183-204.

Siegert, M. J., R. C.A. Hindmarsh, and G. S. Hamilton, 2003. Evidence for a large surface ablation zone in central East Antarctica during the last Ice Age. *Quat. Res*. 59 (2003), 114–121.doi:10.1016/S0033-5894(02)00014-5.

Sinisalo, A., 2007. Geophysical exploration of Antarctic blue ice areas for paleoclimate applications. *PhD thesis*. Arctic Centre Report Series 51.

Sinisalo, A., A. Grinsted and J. C. Moore, 2004. Dynamics of the Scharffenbergbotnen blue-ice area, Dronning Maud Land, Antarctica. *Ann. Glaciol*. 39, p. 417-423.

Sinisalo, A., A. Grinsted, J. C. Moore, H. A. J. Meijer, T. Martma, and R. S. W. van de Wal, 2007. Inferences from stable water isotopes on the Holocene evolution of Scharffenbergbotnen blue ice area, East Antarctica. *J. Glaciol.* 53(182), p. 427-434.

Sinisalo, A., J. Moore, R. van de Wal, R. Bintanja, and S. Jonsson, 2003. A 14-year mass balance record of a blue ice area in Antarctica. *Ann. Glaciol*. 37, p. 213-218.

Sowers, T., R. B. Alley, and J. Jubenville, 2003. Ice core records of atmospheric N_2O covering the last 106,000 years. *Science*, 301, 945-948

Spikes, V. B., 2000. Laser altimetry, mass balance, and meteorites: A two part study of ice streams and blue ice. *M. Sc. Thesis,* The Ohio State University.

Takahashi, S., Y. Ageta, Y. Fujii, and O. Watanabe, 1994. Surface mass balance in east Dronning Maud Land, Antarctica, observed by Japanese Antarctic Research Expeditions, *Ann. Glaciol.*, 20, 242–248.

Takahashi, S., T. Endoh, N. Azuma, and S. Meshida, 1992. Bare ice fields developed in the inland part of Antarctica. *Proc. NIPR Symp. Polar Met. Glaciol.*, 5, 128-139.

Takeuchi, N. 2002. Optical characteristics of cryoconite (surface dust) on glaciers: the relationship between light absorbancy and the organic matter contained in the cryoconite. *Ann. Glaciol*. **34**, 409 – 414.

Taylor, K. C., R. B. Alley, D. A. Meese, M. K. Spencer, E. J. Brook, N. W. Dunbar, R. Finkel, A. J. Gow, A. V. Kurbatov, G. W. Lamorey, P. A. Mayewski, E. Meyerson, K. Nishiizumi and G. A. Zielinski, 2004. Dating the Siple Dome, Antarctica ice core by manual and computer interpretation of annual layering. *J. Glaciol.*, 50(170):453-461.

Van den Broeke, M. R. and R. Bintanja, 1995. The interaction of katabatic wind and the formation of blue ice areas in East Antarctica. *J. Glaciol.*, 41, 395-407.

Van der Kemp, W. J. M., C. Alderliesten, K. van der Borg, A. F. M. de Jong, R. A. N. Lamers, J. Oerlemans, M. Thomassen and R. S. W. van de Wal, 2002. In situ produced 14C by cosmic ray muons in ablating Antarctic ice. *Tellus*, 54B, 186-192.

Van Roijen, J. J., 1996. Determination of ages and specific mass balances from 14C measurements on Antarctic surface ice. *Ph.D. thesis*, Faculty of Physics and Astronomy, Utrecht University.

Van Roijen, J. J., R. Bintanja, K. van der Borg, M.R. van den Broeke, A.F.M. de Jong, and J. Oerlemans, 1994. Dry extraction of $^{14}CO_2$ and ^{14}CO from Antarctic ice. *Nuclear Instruments and Methods in Physics Research, B* 92: 331-334. doi:10.1016/0168-583X(94)96029-1.

Van Roijen, J.J., K. van der Borg, A.F.M. de Jong, and J. Oerlemans, 1995. Ages, ablation and accumulation rates from ^{14}C measurements on Antarctic ice. *Ann. Glaciol.*, 2, 139-143.

Watanabe, O., K. Kamiyama, H. Motoyama, Y. Fujii, M. Igarashi, T. Furukawa, K. Goto-Azuma, T. Saito, S. Kanamori, N. Kanamori, N. Yoshida and R. Uemura, 2003a. General tendencies of stable isotopes and major chemical constituents of the Dome Fuji deep ice core. *Mem. Natl. Inst. Polar Res., Spec. Issue*, 57,1-24.

Watanabe, O., J. Jouzel, S. Johnsen, F. Parrenin, H. Shoji, and N. Yoshida, 2003b. Homogeneous climate variability across East Antarctica over the past three glacial cycles. *Nature* 422: 509-512.

Weller, G. E., 1968. The heat budget and heat transfer processes in Antarctic plateau ice and sea ice. *ANARE Sci. Rep.*, Ser. A(4) Glaciology 102.

Welten, K. C., L. Lindner, K. van der Borg, T. Loeken, L. Schultz, J. Romstedt, and K. Metzler, 1995. Antarctic Meteorites with Unusual Exposure and Terrestrial Histories. *Meteoritics*, vol. 30, no. 5, p. 598.

Welten K. C., K. Nishiizumi, and M. W. Caffee, 2000. Update on terrestrial ages of antarctic meteorites. *Lunar Planet. Sci.* XXXI, CD-ROM 2000.

Whillans, I.M., and W.A. Cassidy, 1983. Catch a falling star; meteorites and old ice. *Science*, 222(4619), 55-57.

Wilch, T.I., W.C. McIntosh, and N.W. Dunbar, 1999. Late quaternary volcanic activity in Marie Byrd Land: Potential 40Ar/39Ar dated time horizons in West Antarctic ice and marine cores. *Geological Society of America Bulletin*, v. 111, p. 1563-1580.

Winther, J.-G.., M. N. Jespersen, and G. E. Liston, 2001. Blue-ice areas in Antarctic derived from NOAA AVHRR satellite data. *J. Glaciol*. 47, 325–334.

Yokoyama, K., 1976. Geomorphological and glaciological survey of the Minami-Yamato nunataks and Kabuto nunatak, East Antarctica. *Ant. Rec.*, 56, 14-19
Yoshida M., H. Ando, K. Omoto, R. Naruse, and Y. Ageta, 1971. Discovery of meteorites near Yamato Mountains, East Antartica. *Antartic Record* 39, 62-65.
Zielinski, G., 1995. Stratospheric loading and optical depth estimates of explosive .volcanism over the last 2100 years derived from the GISP2 Greenland ice core, *J. Geophys. Res.* 100, 20,937-20,955.

Table 1. Details of Antarctic blue ice samples. The sampling sites are indicated in Fig. 1 with the codes given in the first column under the name of the core.

Name	Location	Elev. (m a.s.l.)	Depth/ length (m)	Age (a BP)	Analyses	Reference
SY core (YM)	72°05'S 35°11'E	2150	101 (ver.)	55 000 – 61 000	Gas, chemistry, isotopes	Machida *et al*. 1996, Nakawo *et al*. 1988, Moore *et al.* 2006
SBB01 (SBB)	74°35'S 11°03'W	1173	52 (ver.)	10 500 (+700, -300)	Isotopes	Sinisalo *et al.* 2007
SBB01H (SBB)	74°34'S 11°04'W	1187	100 (hor.)	4426 ±215*	Isotopes	Sinisalo *et al.* 2007
Mt. Moulton (Mt.M)	76° S 135° W	2800	600 (hor.)	10 000 – 150 000	Gas, chemistry, isotopes	Popp *et al.* 2004
MBI#1 (Mt.M)	76°04'S 134°42'W	2820	~30 (ver.)	115 000 – 135 000	Gas, isotopes	Custer 2006

*Age in the middle of the 100-m blue ice core (Sigl and Schwikowski, personal communication). The age span of SBB01H is estimated to be about 500 years (Sinisalo *et al.* 2007).

Figure captions

Fig. 1. Some Antarctic BIAs of scientific interest. SBB=Scharffenbergbotnen, SL="Slöret", YM=Yamato Mountains, Mt.M=Mount Moulton, AH=Allan Hills, FM=Frontier Mountain. Buråsbotnen is located about 50 km northeast from SBB. SBB also indicates the location of Heimefrontfjella mountain range in Dronning Maud Land.

Fig. 2. A BIA in the Scharffenbergbotnen valley surrounded by mountains.

Fig. 3. A sketch of ice flow in a closed type and in an open type (see section 2.5 for definition) BIA. The flow direction is from left to right. Equilibrium line (ELA) separates the snow covered accumulation area from the ablation area i.e. the BIA. The isochrones that represent individual annual layers come up to the surface of the BIA eventually resulting in near-vertical layering in a closed BIA. The oldest ice layers are found at the surface at the end of the BIA while they may never reach the surface on the open BIA.

Fig. 4. Surface characteristics of a BIA. Dust bands can appear as distinct narrow stripes (left) or as more than metre wide bands at the surface of a BIA. Supraglacial moraines and cryoconite holes are visible on the dust band on the right. (Photo: Left: FINNARP/Anna Sinisalo, right: FINNARP/Aslak Grinsted)

Fig. 5. The $\delta^{18}O$ values from the Scharffenbergbotnen valley (left) and "Slöret" (right). The modern day values are measured from snow/firn samples collected in the accumulation areas of the BIAs. Glacial ice was only found in Scharffenbergbotnen showing clearly lower $\delta^{18}O$ values than the modern day samples (Sinisalo *et al.* 2007).

Fig. 6. Supraglacial moraines (black) and nunataks (dark gray) around BIAs (light gray) in Scharffenbergbotnen (left) and around Buråsbotnen, Milorgfjella (modified from Hättestrand & Johansen 2005). See Fig. 1 for locations of the BIAs. The moraines form closed loops in the valley of Scharffenbergbotnen whereas they are drawn out from the cirque in Buråsbotnen.

Fig. 7 A GPR section over a snow ridge between two BIAs in Scharffenbergbotnen valley showing dip angles of isochrones at the firn/blue ice transition zone (see Fig. 1 for location). The main closed-type of BIA is on the left side of the snow ridge, and the smaller open type of BIA on the right side. Ice flows from right to left in the figure. Note that the data are not migrated, i.e. the dip angles of the layers appear steeper than they actually are. (Sinisalo *et al.* 2004)

Fig. 8. Clear annual cycles in the high resolution δD analysis of the horizontal ice core SBB01H from Scharffenbergbotnen BIA (Sinisalo *et al.* 2007).

Fig. 9. Surface age relationship along the flow line of Scharffenbergbotnen. Horizontal age gradient from high resolution isotopic analysis (Fig. 8) of a 101-m long horizontal ice core (Sinisalo *et al.* 2007; gray line), and by dating of internal radar reflection horizons close to the current blue ice/snow transition zone along the flow line (Fig. 7, Sinisalo *et al.* 2004; light gray). The model output was obtained with a linearly changing temporal and spatial surface velocity and accumulation rate reaching the present values in 11 000 years. The model was adjusted to match with the ^{14}C age of a vertical blue ice core (Sinisalo *et al.* 2007). The error

bars for a vertical blue ice core SBB01 are calculated using a radiocarbon calibration curve by Reimer *et al.* (2004). The middle part of the SBB01H was very recently ^{14}C dated to 4426 ± 215 years BP (Michael Sigl, Margit Schwikowski, Paul Scherrer Institut, personal communication). The horizontal distance is measured starting from the bottom of the valley. The SBB01 (Table 1) is located at x=400 m and the SBB01H at x=1400 m.

Fig. 10. Ice thickness H, mass balance b and surface velocity U_s along the SY flow line. The extent of the BIA can be seen from the region of negative b. Bottom: Modelled (Moore *et al.* 2006, Grinsted *et al.* 2003) particle paths (dotted lines) and isochrones (colour contours). The SY core is at 18 km along the flow line. The glacial values of b and U_s optimized by simple scaling to try to match the 60 kyr age of the SY core. Glacial accumulation (b) set to 45% of present day values in accumulation areas (dashed curves), a 5 km reduction in the BIA glacial extent, and glacial flow velocities (U_s) set to 70% of present day values with a delay of 5 kyr between climate shift and ice sheet response, (dashed curves).

Fig. 11. Comparisons of surface age distributions of four BIAs (see Fig. 1 for the locations). Ages for the in the Allan Hills Near Western Ice Field (green), Yamato Mountain (red) and Scharffenbergbotnen (blue) were obtained applying a flow model by Grinsted *et al.* (2003) instead of the oldest ice (28 000 years) in Scharffenbergbotnen that was originally dated by ^{14}C dating (van Roijen 1996). The oldest sample was not found along the main flow line, and its location along the x axis is arbitrary. The snow/blue ice transition is located at 0 at the x axis for the modeled ages. The Mount Moulton data (black) were obtained by dating of tephra bands perpendicular to the ice flow (Dunbar *et al.* 2008, Wilch *et al.* 1999), and it is plotted on x axis so that 0 corresponds to the start of a blue ice transect (Popp *et al.* 2004). The modelled ages are constrained by meteorite terrestrial ages, comprehensive data sets of geophysical and chemical parameters, and radiocarbon ages (for more details see Grinsted *et al.* 2003, Moore *et al.* 2006, Sinisalo *et al.* 2007).

Fig. 1

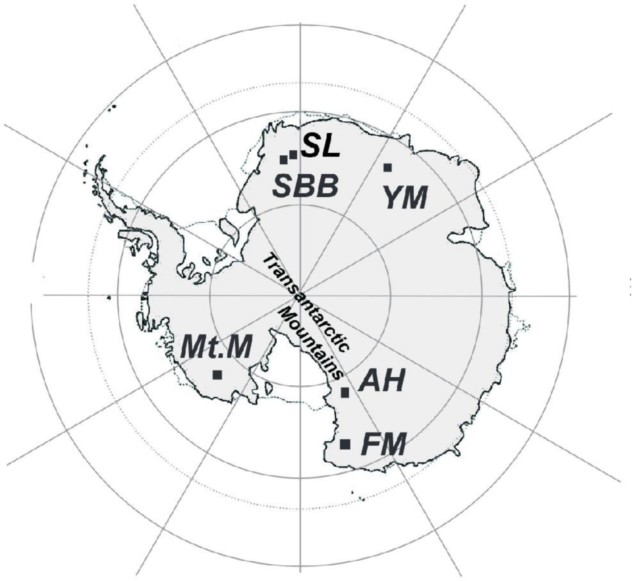
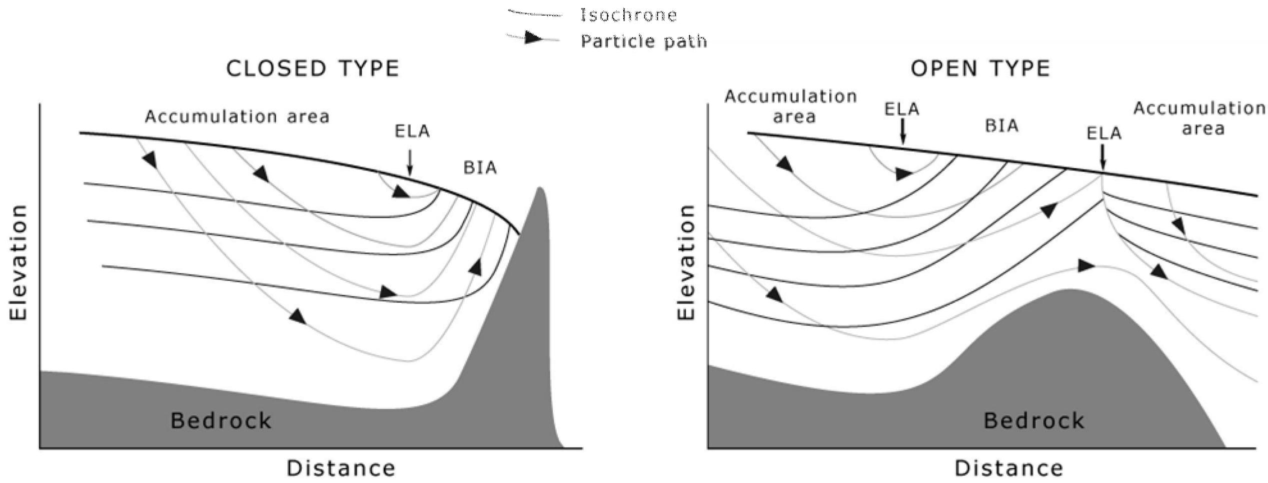


Fig. 2

Fig. 3

Fig. 4



Isochrone
Particle path
CLOSED TYPE
Accumulation area
ELA
BIA
Elevation
Bedrock
Distance
OPEN TYPE
Accumulation area
ELA
BIA
ELA
Accumulation area
Elevation
Bedrock
Distance

Fig. 5

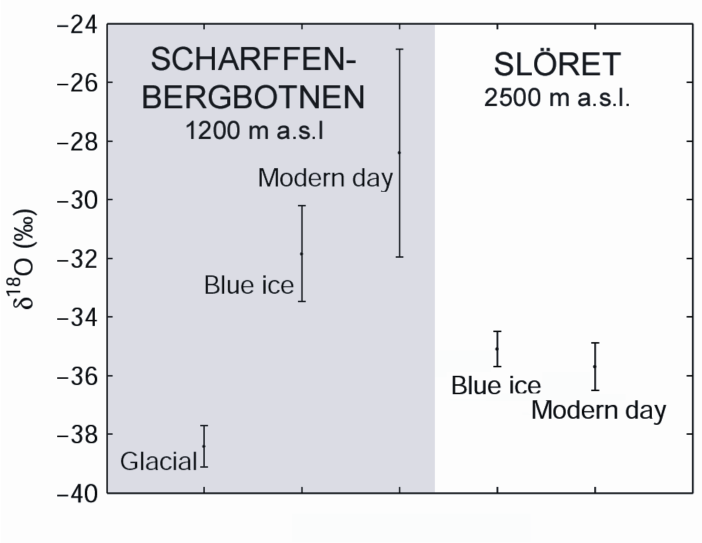

SCHARFFEN-
BERGBOTNEN
1200 m a.s.l
SLÖRET
2500 m a.s.l.
Modern day
Blue ice
Glacial
Blue ice
Modern day
δ18O (‰)
−24
−26
−28
−30
−32
−34
−36
−38
−40

Fig. 6

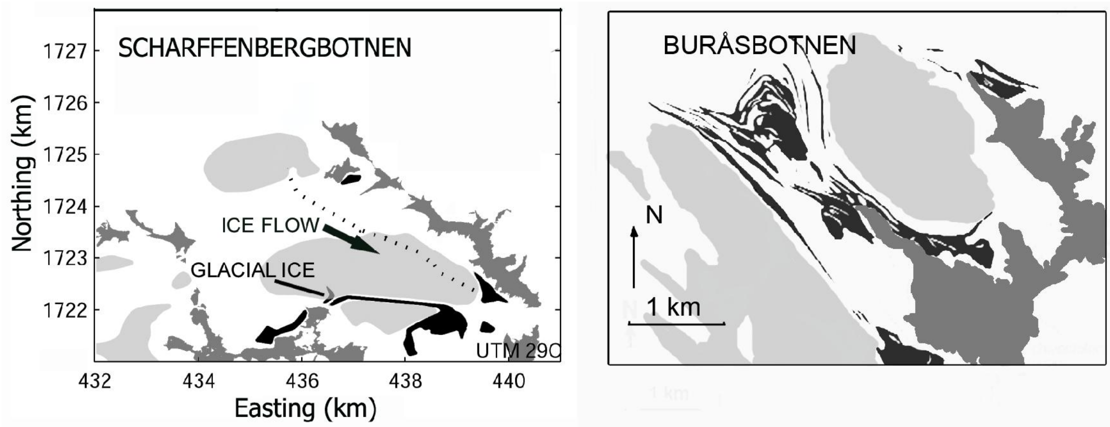
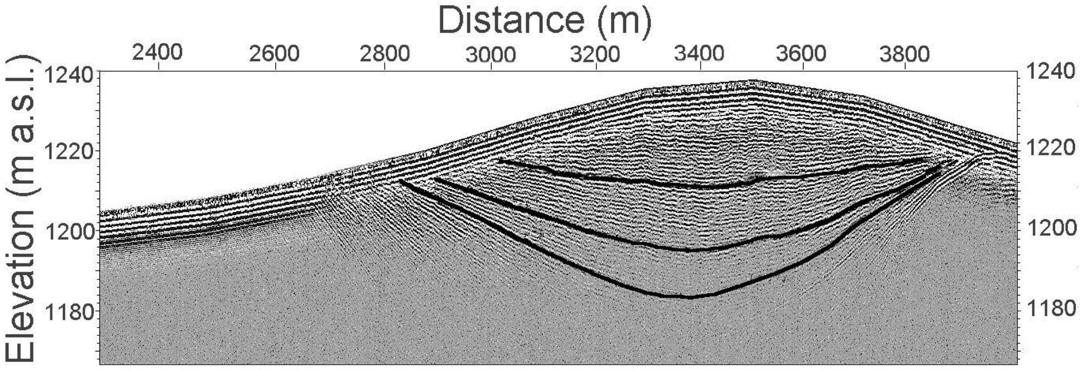

SCHARFFENBERGBOTNEN
1727
1726
1725
1724
1723
1722
Northing (km)
ICE FLOW
GLACIAL ICE
UTM 29C
432
434
436
438
440
Easting (km)
BURÅSBOTNEN
N
1 km

Fig. 7



Distance (m)
2400
2600
2800
3000
3200
3400
3600
3800
Elevation (m a.s.l.)
1240
1220
1200
1180
1240
1220
1200
1180

Fig. 8

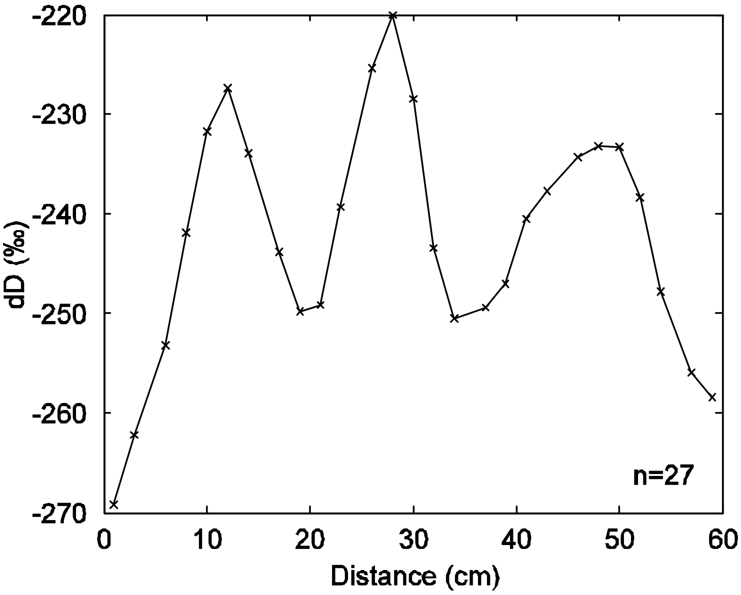


Fig. 9

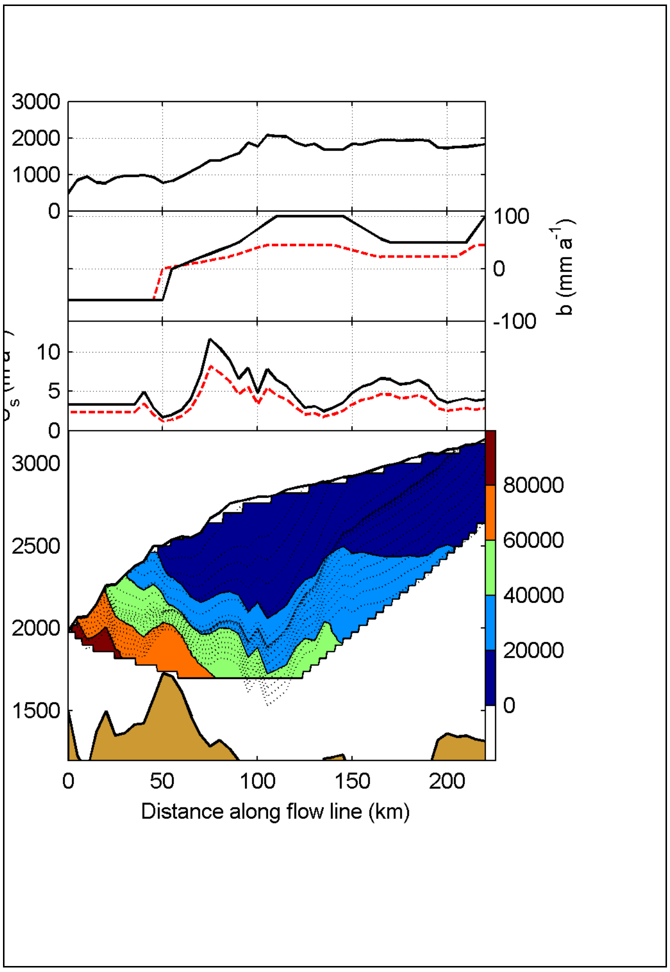
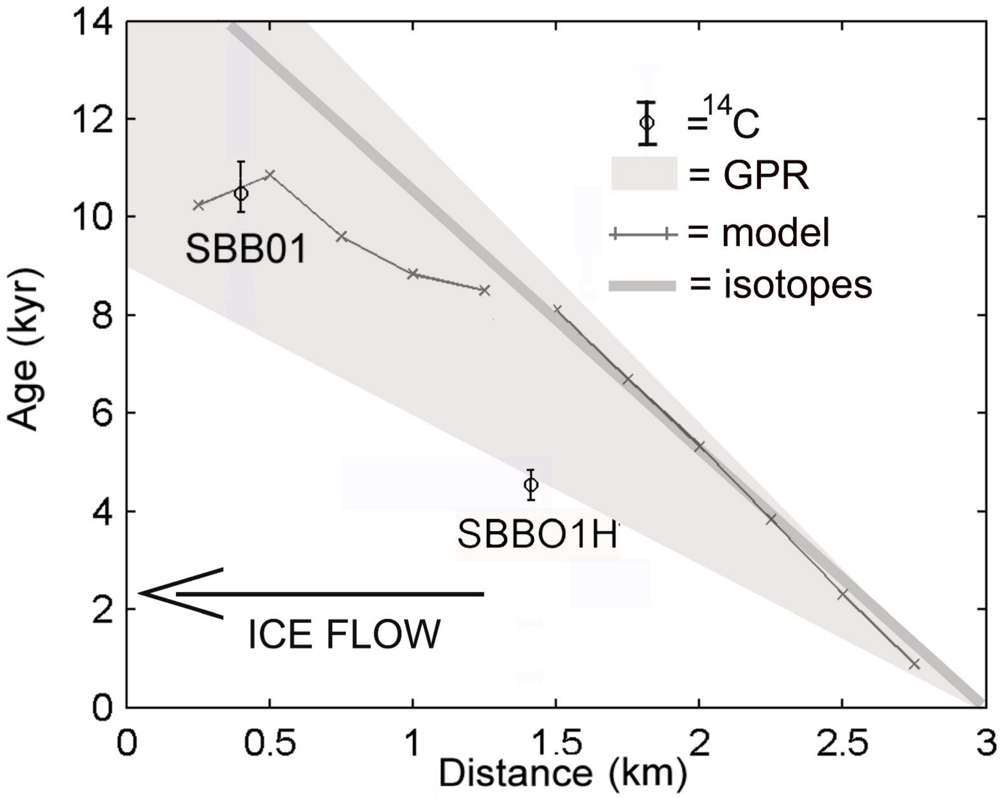

3000
2000
1000
0
100
0
-100
b (mm a^{-1})
10
5
0
3000
2500
2000
1500
80000
60000
40000
20000
0
0
50
100
150
200
Distance along flow line (km)

Fig. 10



= ^{14}C
= GPR
= model
= isotopes
SBB01
SBBO1H
ICE FLOW
Age (kyr)
Distance (km)
14
12
10
8
6
4
2
0
0
0.5
1
1.5
2
2.5
3

Fig. 11