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The two-step shape and timing of the last deglaciation in Antarctica

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Abstract. The two-step character of the last deglaciation is well recognized in Western Europe, in Greenland and in the North Atlantic. For example, in Greenland, a gradual temperature decrease started at the Bölling (B) around 14.5 ky BP, spanned through the Alleröd (A) and was followed by the cold Younger Dryas (YD) event which terminated abruptly around 11.5 ky BP. Recent results suggest that this BA/YD sequence may have extended throughout all the Northern Hemisphere but the evidence of a late transition cooling is still poor for the Southern Hemisphere. Here we present a detailed isotopic record analyzed in a new ice core drilled at Dome B in East Antarctica that fully demonstrates the existence of an Antarctic cold reversal (ACR). These results suggest that the two-step shape of the last deglaciation has a worldwide character but they also point to noticeable interhemispheric differences. Thus, the coldest part of the ACR, which shows a temperature drop about three times weaker than that recorded during the YD in Greenland, may have preceded the YD. Antarctica did not experienced abrupt changes and the two warming periods started there before they started in Greenland. The links between Southern and Northern Hemisphere climates throughout this period are discussed in the light of additional information derived from the Antarctic dust record.

Introduction

Since the very well documented and thorough review of Rind et al. (1986) published in the first issue of Climate Dynamics, numerous studies have been dedicated to the Younger Dryas (YD), a climatic stage which took place during the second half of the last deglaciation. This cold event, originally defined for a pollen zone in Europe (Jensen 1938; Iversen 1954), followed the warmer Bölling and Alleröd periods and spanned approximately a millenium from ~11 to 10 ky ^{14}C ages BP. New information obtained in recent years concern various aspects of this BA/YD climatic sequence. Of particular importance and directly relevant to the present study are those relative to the absolute dating of this sequence, to climate mechanisms potentially involved, to the abruptness of associated climatic transitions (at least in the North Atlantic and adjacent regions where this event is better defined) and to its possible worldwide geographical extent.

Dating of submerged corals contemporaneous with the last deglaciation from Barbados and Mururoa by the uranium/thorium method (Bard et al. 1990a) allowed Bard et al. (1993) to propose a simple linear calibration of the conventional ^{14}C ages for the interval between 8 and 20 ky BP[1]. Accounting for uncertainties, this calibration places the termination of the Younger Dryas from 11.5 to 11.2 ky BP (Bard et al. 1993). This absolute dating is independently confirmed by chronologies obtained by counting annual layers on the GRIP (Johnsen et al. 1992) and GISP 2 (Alley et al. 1993b) Central Greenland ice cores in which this termination is dated respectively at 11550±70 and 11640±250 y BP. It also agrees with varve calibration performed in Lake Gosciaz (Rozanski et al. 1992) in Central Europe (Poland) and with tree-ring calibration of the ^{14}C method (Kromer and Becker 1992). There are still some dating differences in the concurrent change in other European records either attributed to ^{14}C dating calibration (Hadjas et al. 1993) or to a real lag of climatic change between Greenland and Central Europe (Becker and Kromer 1993). However, as summarized by Alley et al. (1993b), one can confidently accept the age of 11.5±.2 ky BP for the end of the YD event. As seen in the GRIP and GISP 2 cores, this event lasted more than a millenium (respective estimates of 1150 and 1300 y).

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[1] All the ages given in this article are calendar ages (ky before present). When using ^{14}C ages, the calendar ages given in the text are calculated using the formula proposed in Bard et al. (1993): age (ky BP) = 1.24 (age ^{14}C-.84)

Because they permit changes in climate or climate parameters to be observed at subdecadal or even annual time scales, ice cores provide a more accurate estimate of how rapidly such changes occurred during the climatic transition associated with the BA/YD sequence. Dansgaard et al. (1989) showed that the Younger Dryas/Preboreal transition was very abrupt in the South Greenland Dye 3 core with temperature increases of $\sim 7°$ C in a few decades and even more rapid atmospheric circulation changes. This is fully confirmed in Central Greenland both for temperature changes (Johnsen et al. 1992; Grootes et al. 1993) and for rapid reorganization (<5–20 y) in atmospheric circulation (Taylor et al. 1993; Mayewski et al. 1993). There are in these records indications of an even more rapid change in precipitation pattern with snow accumulation possibly doubling in one to three years (Alley et al. 1993b). The transition from the Older Dryas to the Bölling/Alleröd that occurred around 14.5 ky BP (Johnsen et al. 1992; Alley et al. 1993b) was also very abrupt. As shown in the GRIP isotopic record (Fig. 4), the cooling phase leading to the Younger Dryas is gradual compared to these abrupt warmings, giving the BA/YD sequence an asymetric sawtooth shape (Johnsen et al. 1992).

That there exists a strong link between the BA/YD sequence and changes in the North Atlantic oceanic circulation was already discussed in the Rind et al.'s (1986) review. Broecker et al. (1985) previously suggested that this sequence as well as the rapid "Dansgaard-Oeschger" oscillations found during the late glacial, are caused by fluctuations in the rate of formation of deep water in the northern Atlantic that were already documented (Ruddiman and McIntyre 1981). A considerable amount of new information about this link including the routing of meltwater from the Laurentide ice sheet (Broecker et al. 1989), the existence of two meltwater pulses (see Fig. 4) at the end of the deglaciation (Fairbanks 1989) and indications of oceanic circulation changes (Boyle and Keigwin 1987; Jansen and Veum 1990; Lehman and Keigwin 1992; Duplessy et al. 1992) are now available. In parallel, the Younger Dryas event has motivated numerous modeling studies ranging from relatively simple conceptual models (Birchfield 1989; Broecker et al. 1990; Birchfield and Broecker 1990; Stocker and Wright 1991) to numerical experiments using atmospheric and oceanic general circulation models (Rind et al. 1986; Maier-Reimer and Mikolajewicz 1989). They also include models of intermediate complexity (Wright and Stocker 1993). Despite this combined effort based on new data and dedicated modeling, it is not yet fully understood how various mechanisms affected the ocean/atmosphere system in the North Atlantic during the BA/YD sequence (Broecker 1989; Fairbanks 1990; Jones 1991b; Zahn 1992; Veum et al. 1992).

Beyond these studies which clearly confirm that the BA/YD sequence is quite well characterized in Greenland and in the North Atlantic and the search for mechanisms linked to North Atlantic deep water formation, numerous studies showed that the Younger Dryas or "Younger Dryas type" events have a larger geographical extent than initially thought and could even be global in character (Alley et al. 1993a). For the Northern Hemisphere, such records include those obtained in the Caribean and in the Gulf of Mexico (Overpeck et al. 1989; Flower and Kennet 1990), Africa (Street-Perrott and Perrott 1990; Gasse et al. 1989; Roberts et al. 1993), western (Engstrom et al. 1990) and eastern (Levesque et al. 1993) North America, China (Gasse et al. 1991; An et al. 1993), the Gulf of California (Keigwin and Jones 1990), the western North (Kallel et al. 1988) and tropical (Linsley and Thunnell 1990) Pacific and in the Sulu Sea in the eastern Indonesian Archipelago (Kudrass et al. 1991). (Note however that the interpretation of some of these records is questioned in the recent review of Alley et al. 1993a).

For the Southern Hemisphere, the evidence of a "Younger Dryas type" event is less well documented. Beyond the records cited in Rind et al. (1986) for South America, South Africa, South Georgia and New Zealand, there are new indications for New Zealand by Newnham (unpublished manuscript) and more recently by Denton and Hendy (1994), for the Southern Ocean (Labracherie et al. 1989) and north of Australia (de Deckker et al. 1991). There is a controversy in the interpretation of southernmost South America data (Markgraf 1991, 1993); whereas Heusser and Rabassa (1987) and Heusser (1989) interpret those pollen records as reflecting cooling comparable to the Younger Dryas, Markgraf (1993) suggests instead that observed changes are a response to local and regional disturbance by fires.

Although the worldwide character of the Younger Dryas is yet to be fully demonstrated, this possibility can now be taken seriously. However the timing of the "cold event" recorded in various parts of the world at the end of the last deglaciation with respect to the well-dated North Atlantic cooling still has to be firmly established. In this context, the Antarctic record of the deglaciation is of obvious interest. In addition, the fact that ice cores offer a unique opportunity to obtain high resolution records of many parameters relevant to climate and environmental changes gives a reasonable hope that such studies may help to understand how various parts of the climate system were linked during this period. With this in mind, we have already discussed the Dome C (Lorius et al. 1979) and Vostok (Jouzel et al. 1987b) records. They indicate that the climatic transition in East Antarctica was a two-step process with two warming trends interrupted by a cold reversal (Jouzel et al. 1987a; Jouzel et al. 1992).

Here, we examine further the link between this Antarctic cold reversal (ACR) and the BA/YD sequence and more generally between the Southern Hemisphere and Northern Hemisphere climates during the deglaciation. This is interesting mostly for two reasons: (1) we now have new datasets obtained along a 780 m core drilled at the Dome B site (Fig. 1) covering $\sim$ the last 30 ka which allows the deglaciation to be quite well documented in East Antarctica both for cli-

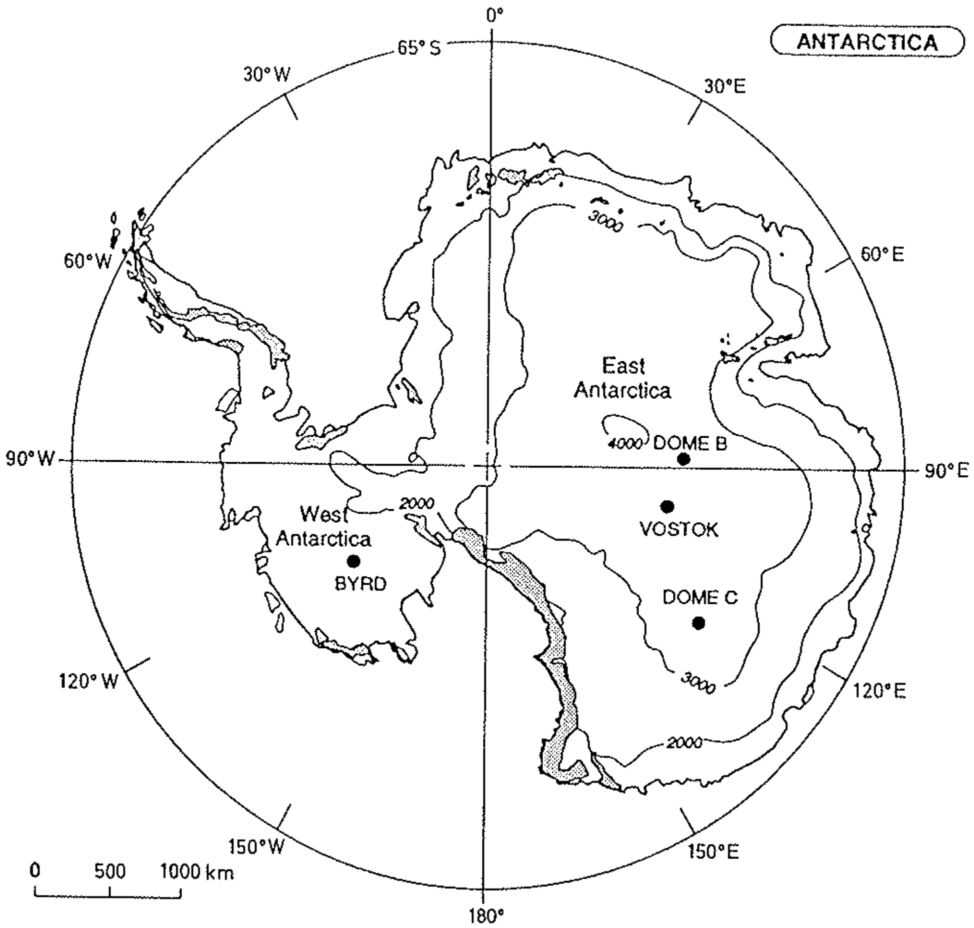

Fig. 1. Map of Antarctica with Dome B and other Antarctic sites

mate and dust records and (2) a large part of the earlier dating problems existing both for the BA/YD (see earlier discussion) and the ACR events (Jouzel et al. 1987a; Jouzel et al. 1992) are or may now be solved.

The Dome B ice core

Records from four deep Antarctic ice cores are examined in this study. For three of these sites, Byrd, Dome C and Vostok, the discussion is based on published data (Johnsen et al. 1972; Lorius et al. 1979; Jouzel et al. 1989; Jouzel et al. 1992; Hammer et al. 1994) and we focus this section on the presentation of the new records obtained from Dome B core. Dome B is a very cold and relatively low accumulation site located on the high East Antarctic plateau 320 km upflow from Vostok and 870 km from Dome C (Fig. 1). A thermal drilling was conducted during the 1987–1988 Austral season by the 33th Soviet Antarctic Expedition. Two series of samples (2 m and 1 m ice increments) were cut in the field and analyzed for $\delta^{18}O$ (in Tallin and Saclay respectively). The deuterium profile was analyzed on the 1 m series and is given in Fig. 2 along with the dust concentration measured by Coulter counter. In addition, we have undertaken a very detailed isotopic study, i.e., with 10 samples per meter, for the deglaciation part of the record on which we focus the present discussion. This sampling permits a subdecadal resolution and thus a thorough examination of the rapidity of climatic change.

For this discussion, we use two parameters, δD, a proxy of temperature change over Antarctica and dust concentration, an indicator of changes in dust source strength and atmospheric circulation. We note two striking features for the deglaciation part of the records. First, deglaciation is interrupted around 480 m by a well-marked return to colder conditions (more negative δD). Second, the large dust fallout level during the Last Glacial Maximum is followed by a return to low Holocene values which occurred just before the end of the first warming period (around 490 m).

As for other East Antarctic cores (Lorius et al. 1979; Lorius et al. 1985; Ritz submitted), accurate dating of Dome B is difficult because low accumulation prevents isotopic or chemical seasonal variations to be used for annual counting. Dating may be obtained by combining a simple one-dimensional vertical thinning model and an accumulation model which accounts for lower accumulation during colder periods (Ritz submitted; Robin 1977). Either an estimate of modern accumulation (Jouzel et al. 1989) or an independent estimate of the age at some given depth (Jouzel et al. 1993) is needed as additional input.

To constrain the Dome B chronology, we first propose to use the information contained in the changes of terrestrial and marine aerosol fallouts over Antarctica during the first part of the deglaciation. Both for

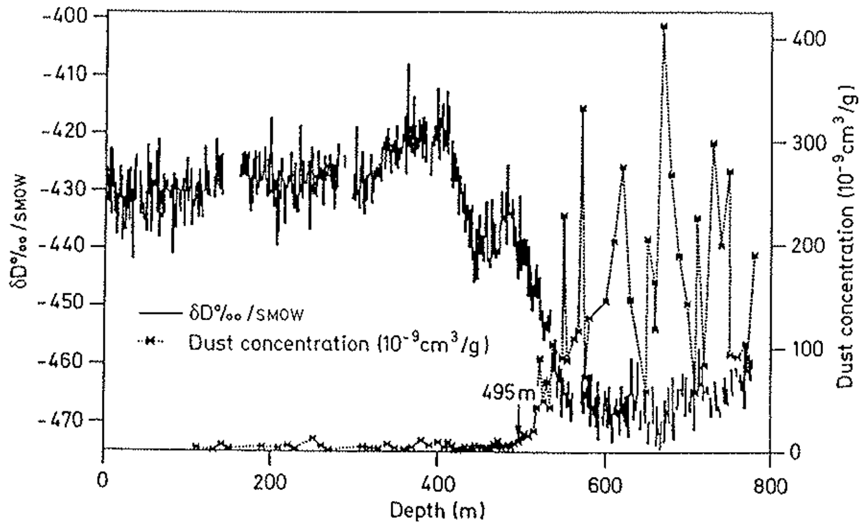

Fig. 2. Depth profiles of the deuterium and of the dust contents at Dome B with indication of the end of the dust peak

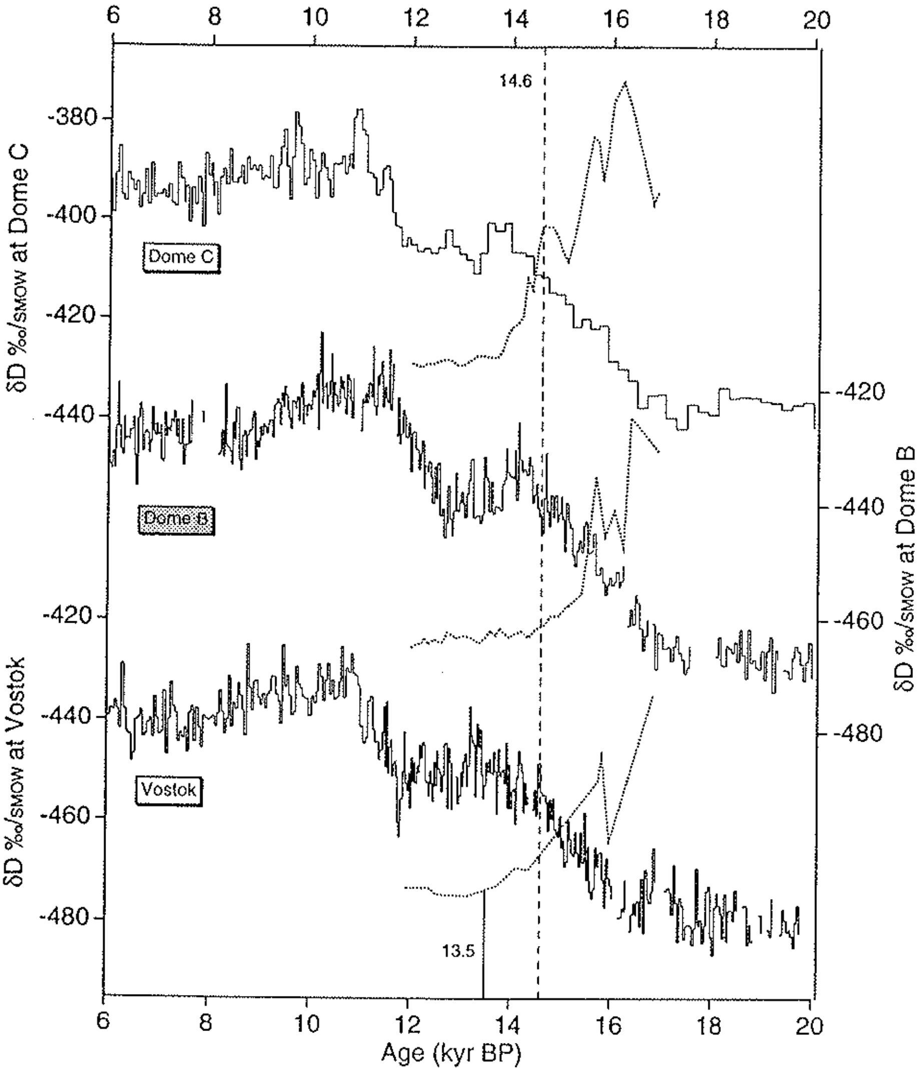

Fig. 3. Deuterium and dust profiles against age for Dome B, Dome C and Vostok. Time scales are from Jouzel et al. (1989) for Dome C and from Jouzel et al. (1993) for Vostok. The Dome B time scale is discussed in the text. The dust profiles (*dotted line*, arbitrary scale) are given only for the deglaciation part. The *vertical line* (14.6 ky BP) corresponds to the end of the dust peak at Dome B

dust and sodium concentration, there is a strong decrease between the peak characteristic of the very cold conditions prevailing during the Last Glacial Maximum and the end of the first warming trend where fallout is quite similar to the Holocene values. These changes are seen both in the East Antarctic cores (see Fig. 3 for Vostok, Dome C and Dome B), and in the Byrd core in West Antarctica. As reviewed in Vaikmae et al. (in preparation) various indicators measured in this Byrd core (Cragin et al. 1977; Legrand 1985) indicate that aerosol fallouts reached Holocene values at 1175±25 m.

The Byrd core has the advantage of being more accurately dated than the East Antarctic cores. Hammer et al. (1994) used seasonal cycles that are preserved in this core whereas Sowers et al. (1992) derived a chronology in correlating the Byrd record of $\delta^{18}O$ of O_2 in air with that measured in the GISP2 Greenland record. At a depth of 1175 m, the age of the Byrd core given by the two chronologies are respectively 14.6 ky BP (Sowers et al. 1992) and 14.1 ky BP (Hammer et al. 1994). These two figures agree within dating uncertainties estimated to be ±5% in Hammer et al. (1994). In the Sowers et al. (1992) chronology they include the uncertainty on the GISP time scale itself, 0.4 ky at this level (Alley et al. 1993b), and those due to the differences between the age of the air and the age of the ice at GISP2 and Byrd. At Dome B, the corresponding Holocene dust level is reached at a depth of 495±5 m. Although dust sources are quite possibly different in East and West Antarctica, we can confidently use this level as a stratigraphic time marker because major aerosol changes are governed by the same processes (changes in sea level, extension of arid areas and atmospheric circulation). Indeed, Delmas and Petit (1994) recently suggested from a close examination of the dust chemical composition that there is a nearly common source of ice age Antarctic aerosol over the whole continent.

The accuracy of the derived Dome B chronology (Fig. 3) is no better than ±1 ky at this level largely because of the coarseness of the Byrd aerosol records which presently constitutes one of the main source of uncertainty. However, comparison of the Byrd continuous isotopic record with that of Dome B (Fig. 5) gives additional weight to the proposed correlation: the end of the warming trend is quite synchronous in Byrd and Dome B once the aerosol fallout is used as a time marker (as discussed below the subsequent cooling is less well marked in the Byrd than in the East Antarctic records).

We opted to present and discuss the Dome B isotopic profile in using a time scale in which we assign the age of 14.6 ky BP for the depth of 495 m corresponding to the return to Holocene dust levels. We favor this choice based on the age given by Sowers et al. (1992) at the corresponding Byrd level because it is supported by independent arguments; first (see later) from the comparison between the Greenland and Antarctic methane records, second, from comparison with deep-sea core records which is examined in the following section.

Vostok and Dome C records (deuterium and dust) are presented in Fig. 3, along with those of Dome B, using the recently published EGT chonology (Jouzel et al. 1993) for Vostok and that derived by Jouzel et al. (1989) for Dome C. The end of the Vostok dust peak is reached at 13.5 ky BP apparently later than discussed for Dome B. However trace gas measurements (Sowers et al. 1993; Chappellaz et al. 1993) have suggested that the Vostok EGT time scale is too young for this period. Using the well-marked Younger Dryas GRIP methane drop, also recorded in Vostok (Chappellaz et al. 1990), as an interhemispheric time marker requires a shift in the Vostok age by 1.2 ka (Chappellaz et al. 1993). This would put the end of the Vostok dust peak very close to 14.6 ky BP. We note however that this argument in favor of our Dome B chronology suffers from the uncertainty in the determination of the air age/ice age difference (this uncertainty associated with the close-off of air bubbles is higher in low accumulation sites such as Vostok than in Byrd where the accumulation is higher). Obviously comparing the Greenland data to a record of concentration in methane at Dome B would be very useful. Such a record is not available because the Dome B ice is not of sufficiently good quality for performing reliable measurements on air bubbles.

Obviously, both the end of the dust peak and of the warming trend must be synchronous in the three East Antarctic cores which is not the case with the chronologies we have chosen to present in Fig. 3. This would require slight adjustments of the order of 1 ky (between Dome B and Vostok) and slightly less between Dome B and Dome C. As none of the three chronologies is known with an accuracy better than 1 ky, we do not yet think it justified to propose any revision of the Vostok and Dome C published chronologies. We focus the following discussion on comparisons between Dome B and other records but our conclusions will remained unchanged if either Vostok or Dome C records were used once for these slight adjustments were made.

Comparison with deep-sea core records

Indirect indication of the validity of the dating of the end of the dust peak recorded in the Antarctic ice cores may be found in deep-sea cores. This approach is based on the interpretation of the ice core dust record along with that of deep-sea records indicative of changes in North Atlantic Deep Water (NADW) production and in sea ice extent around Antarctica. Isotopic studies indicate that Last Glacial Maximum dust found in Dome C largely originated from South America (Grousset et al. 1992). The main potential contributor is the very broad Argentine continental shelf (Grousset et al. 1992) which extended ~1000 km from the present coast line when sea level was lowered by 110 m. The subsequent deglaciation is marked by two intervals of massive ice-sheet melting (Fairbanks 1989, 1990). The first meltwater pulse (MWP1a), centered around 14.1 ky BP, contributed to a fast sea-level rise of 30 to 50 m and to the submergence of about 65% of the exposed Argentine continental shelf largely south of 44°S (Burckle et al. in preparation) cutting off a major source of dust to East Antarctica. Virtually coincident with this meltwater pulse, the formation of NADW was re-initiated between 14.1 and 14.6 ky BP (Fig. 4) as deduced from the $\delta^{13}C$ study of a southern Atlantic deep-sea core (Charles and Fairbanks 1992).

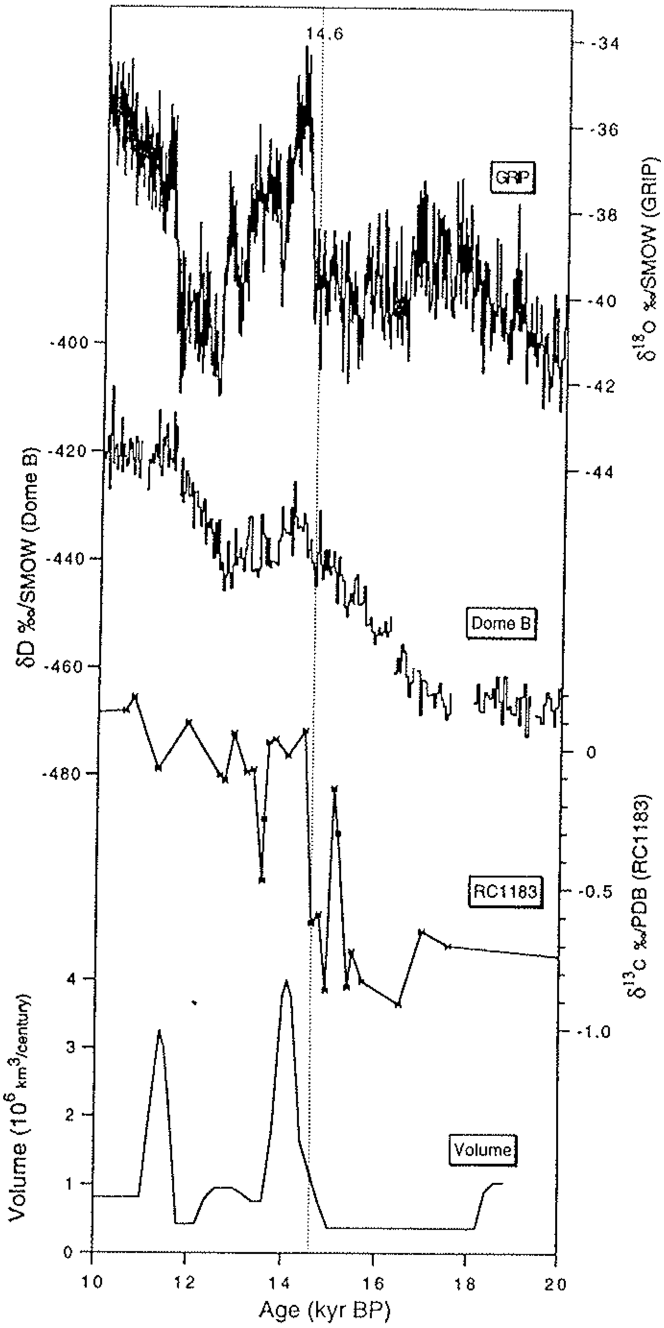

Fig. 4. Age profiles of (from top to the bottom), the $\delta^{18}O$ GRIP profile (adapted from Johnsen et al. 1992), the Dome B profile (this work), the $\delta^{13}C$ profile in the RC11-83 South Atlantic deep-sea core (adapted from Charles and Fairbanks 1992) and the deglaciation rate (adapted from Fairbanks 1989 and from Bard et al. 1993). The *vertical line* (14.6 ky BP) corresponds to the end of the dust peak at Dome B

At the same time the relative abundance record of the sea-ice related diatom, E. Antarctica, shows a drastic decrease indicative of a southward retreat of the northern edge of the sea-ice (Burckle et al. in preparation). This apparent synchronism, attributed to the heat release related to NADW activity, supports the view that, when NADW production rates are high, Southern Hemisphere temperatures increase rather than the opposite (Crowley 1992). This well-documented sea-ice retreat was probably associated with a decrease in the polar front activity which could further help to reduce aridity and dust mobilization in southern South America (Burckle et al. in preparation). Thus, we have through the conjunction of these various processes a qualitatively reasonable explanation for the drastic decrease of Antarctic dust fallout during the warming trend. The chronology of these oceanic events, sea level change resulting from MWP1a and aridity change associated with sea-ice retreat due to the NADW reinitiation, is roughly consistent with the dating of the end of the dust peak at ~14.6 ky BP.

Comparison with the Greenland records

We first note that there is a major dust fallout in Greenland cores with a return to low values dated around 14.4 ky BP and 14.7 ky BP in the GRIP (GRIP project members 1993) and GISP2 (Taylor et al. 1993; Mayewski et al. 1993) cores respectively, i.e., just at the beginning of the Bölling warm period. However, that this major dust drop apparently occurred synchronously in Antarctica and Greenland may be fortuitous because the atmospheric residence time of dust is only a few weeks and thus too short to allow intense mixing in the global atmosphere. There are indeed significant differences between the Antarctic and Greenland dust record: (1) whereas the dust drop is slow in Antarctica it is quite rapid in Greenland which can only be caused by changes in atmospheric circulation and (2) unlike in Greenland where dust levels are again relatively high in the YD, the Antarctic dust levels remain low, i.e., at their Holocene value, through the ACR.

GRIP and Dome B isotopic profiles are compared in Fig. 4. Following the temperature interpretation adopted both for Antarctic (Jouzel et al. 1993) and Greenland cores (Johnsen et al. 1992) the BA/YD surface temperature drop is ~10°C (6.5‰ in $\delta^{18}O$) whereas the strength of the ACR (~20‰ in δD) is estimated to be around 3°C at the surface level (and 2°C above the inversion). This temperature drop probably represents the East Antarctic Plateau well as a whole. The Vostok and Dome B deuterium decreases are quite similar and the lower Dome C value is likely due to the coarse resolution of this record (Jouzel et al. 1992).

We now examine how this factor of ~3 between the temperature drops at GRIP and in East Antarctic cores is altered if the more detailed Dome B record is used. We have analyzed the Dome B core on 10 cm samples in such a way as to get a time resolution similar to the one which will be ultimately reached in the GRIP core where the deglaciation is now being studied with a depth resolution of ~14 cm. Over the 15 to 11 ky BP time interval, the average annual layer thickness is 2.8 mm at Dome B and 4.5 mm at GRIP leading to an average time resolution of about 3 years in each of the records. In Fig. 4 we use the GRIP curve based on 55 cm samples (Johnsen et al. 1992). Similar time reso-

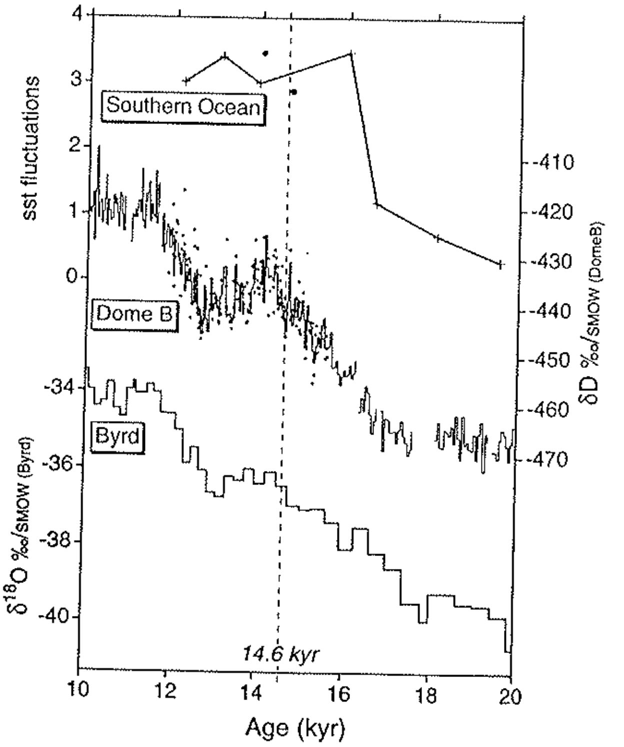

Fig. 5. Ages profiles (from top to bottom) sea surface temperature (diatoms) in Indian Ocean core MD551 adapted from Bard et al. (1990b); the Dome B deuterium profile: the *line* represents the 1 m record (as Fig. 4), the *dots* correspond to 40 cm intervals (see text) and the Byrd oxygen 18 record adapted from Hammer et al. (in press). The *vertical line* at 14.6 ky BP corresponds to the major aerosol drop taken as a time marker

lution corresponds to 30/40 cm samples at Dome B. Thus, we compare in Fig. 5 the Dome B curve of Fig. 4 constructed with 1 m samples and that, still discontinuous, obtained in regrouping four successive 10 cm samples. This clearly adds variability to the record. This variability becomes much larger with 10 cm samples (not shown) and may partly be due to isotopic noise resulting from deposition processes. It does not however significantly change the total amplitude of the Antarctic cold reversal.

The Dome B/GRIP comparison (Fig. 4) illustrates that the shape of the deglaciation is significantly different in Greenland and Antarctica. The coldest part of the ACR, between ~13.5 and 12.5 ky BP appears to precede the YD by ~1 ka. Such a lead has been previously found from the study of oceanic records in the Indian sector of the Southern Ocean (Labracherie et al. 1989) which further supports our chronology as Antarctica and the Southern Ocean are expected to be in phase. On the other hand, Denton and Hendy (1994) recently documented a glacier advance in the Southern Alps of New Zealand that was coeval with the Younger Dryas suggesting that this Southern Hemisphere lead may have been limited to Antarctica and to the Southern Ocean.

These differences between Greenland and Antarctica also concern the period of warmings. Thus, the Antarctic temperature starts to rise around 12.5 ky BP before the end of the Younger Dryas. The situation is somewhat similar for the first warming trend which, at Dome B and other Antarctic cores, starts around 17 ky BP, and, apparently, even earlier at Byrd (this difference most probably results from dating uncertainties: the Vostok-SPECMAP correlation performed by Sowers et al. (1993) suggests that the EGT Vostok time scale should be too young by up to 4 ky during this period). In Greenland, temperatures steadily increase from 22 to 16.5 ky BP but then return to slightly colder conditions before the main warming trend which at 14.5 ky BP leads to the Bölling period. This Greenland warming which occurs in only about century takes place well after Antarctic temperatures started to increase. Here, again, the Southern Hemisphere lead is documented in the Southern Ocean as illustrated in Fig. 4 from an oceanic record located at 55°S (Bard et al. 1990b) and fully discussed in Imbrie et al. (1992). Another striking feature is that unlike in Greenland, and more generally in the North Atlantic (Bard et al. 1987; Lehman and Keigwin 1992), the warming periods are quite smooth in Antarctica and do not correspond to rapid climatic changes. This is true both for the first warming trend and for the transition to the Holocene (rates of temperature changes do not appear modified when the high resolution Dome B record is examined). The warmest temperatures (climatic optimum) are already reached around 11.5 ky BP in Dome B whereas full Holocene climate is reached ~1 ka later in central Greenland. The Dome B data thus confirm the existence of an early climatic optimum in East Antarctica (Ciais et al. 1992).

Figure 5 indicates that there is no such strong temperature reversal in the Byrd record. This point, fully discussed in Hammer et al. (1994), illustrates a different situation in East and West Antarctica possibly because West Antarctica is less influenced by changes in the oceanic conveyor belt than East Antarctica. From visual inspection we can however associate relatively low isotopic values observed in Byrd around 13 ky BP to the ACR.

Discussion

This comparison of the Antarctic and Greenland climatic records is indicative of the complexity of the links between Northern and Southern Hemisphere climates during the last deglaciation (Imbrie et al. 1992). We have put arguments together that support the key rôle of NADW formation through its influence on Southern Hemisphere sea-ice extent. However, it appears that the Southern Ocean $\delta^{13}C$ record, (Charles and Fairbanks 1992 and Fig. 4), and thus NADW formation shows little change during the Younger Dryas (see also Jansen and Veum 1990) and we have no clear indication that the ACR cooling is related to oceanic circulation change. Given available data, it is difficult to invoke, as has been suggested (Kudrass et al. 1991), that changes in greenhouse gases may explain a climatic link between both hemispheres at this time. In particular, the direct radiative impact of changes in methane (Chappellaz et al. 1990, 1993) would not account for more than 0.1° C (Chappellaz et al. 1990). Also, the fact that warming was felt in Antarctic cores well before it was observed in Greenland support the idea that the interhemispheric link operated differently during the initial stage of the Southern Hemisphere deglaciation.

Imbrie et al. (1992) thoroughly discussed the respective timing of the last deglaciation between the Northern and Southern Hemispheres in the context of the Milankovitch theory which indicates that the intitial response to initial forcing must occur in the Northern Hemisphere. The authors noted that early Northern Hemisphere deglaciation is supported by the change in the position of the southern margin of the Laurentide ice sheet (Broecker and Denton 1989) and by oceanic records from the Greenland Sea (Jones 1991a). To explain the early Southern Hemisphere response, with respect to the main deglaciation step associated with the melting of the Laurentide ice sheet, it is postulated that the initial response to the insolation forcing which involves the Nordic Sea is rapidly transmitted to the Southern Hemisphere (Imbrie et al. 1992; Lehman et al. 1991).

The timing and the shape of the deglaciation as seen from Dome B and other Antarctic records do not contradict, and indeed fits relatively well in the sequence of changes described in Imbrie et al. (1992). As for the sea surface temperature at 55°S, the Dome B first warming trend slightly lags the Nordic deglaciation (see Fig. 1 of Imbrie et al. 1992). The smoother shape of the deglaciation in the Southern Hemisphere would then result from the larger inertia of the climate response in this hemisphere. However, our results do not fundamentally add to the conclusion about the respective behavior of the Northern and Southern hemispheres during the last deglaciation which, as noted by Imbrie et al. (1992), has still to be proven.

In this context, we would like to raise the two following issues which both aim to emphasize that conditions prevailing in the Southern Hemisphere may have an important role during this climatic period, i.e., that not all is driven by Northern Hemisphere insolation changes or by processes taking place in the North Atlantic. First, as suggested by Crowley and Kim (1994) for the Last Interglacial, it may be too narrow a view to limit the Milankovitch climate connection just to the field of insolation forcing at 65°N. Because it is thought to play an important role in the sea-ice extent around Antarctica, we previously examined how summer insolation at 60°S may explain part of the Vostok temperature change (Genthon et al. 1987). In this line, determining the geographic response to total insolation changes as was done for the Last Interglacial (Crowley and Kim 1994) would be certainly worthwhile for better understanding the sequence of changes during the last deglaciation. Second, and this is already discussed in Imbrie et al. (1992), what happens in the Southern Ocean may be of importance for the Atlantic's overturning and thus for the formation of NADW. Using a global ocean model, Toggweiller and Samuels (1992) suggested that the magnitude of the deep outflow from the Atlantic is proportional to the westerly wind stress at the latitude of the tip of South America. The fact that the major dust drop in Antarctica, which quite probably results from large atmospheric circulation changes, is roughly synchronous (~14.6 ky BP) with the major change in NADW reflected in $\delta^{13}C$ RC1183 record (Fig. 4) may be seen as an argument in favor of the Toggweiller and Samuels suggestion.

We admit that these considerations about the role of Southern Hemisphere are somewhat speculative at the present stage of understanding of ocean circulation changes. Even if we feel compelled to discuss our Antarctic data in the framework of the deglaciation theory developed by J. Imbrie and his coworkers (Imbrie et al. 1992), we think that the main value of our study stems from the data themselves. We hope we have now, through the new Dome B datasets and earlier Byrd, Dome C and Vostok records, a more comprehensive view of the timing and shape of the last deglaciation in Antarctica. The data presented here definitively show that the two-step shape of the last deglaciation has a worldwide character with a less accentuated return to cold conditions in Antarctica than in Greenland and in the North Atlantic regions. Despite the lack of accurate absolute dating, the use of stratigraphic markers allows us to suggest such interesting chronological informations as the fact that at least the coldest part of the Antarctic cold reversal may have preceded the Younger Dryas by ~1 ka and the two warming periods started there before they started in Greenland.

We obviously recognize the need for a more accurate chronology of the last deglaciation in East Antarctica. There is a hope of finding a sufficiently high accumulation site in its Atlantic sector to allow either absolute dating by varve counting or, at least, accurate comparison with the Greenland records in using various changes in the composition of the atmosphere as global stratigraphic markers. Deep drilling is planned in this sector in the framework of the European Project for Ice Coring in Antarctica (EPICA). This may never be possible in the central part of East Antarctica where accumulation is too low. We feel that the strategy we have followed here, which consists in correlating central East Antarctic records to well dated records from West Antarctica in using changes in dust concentration as a stratigraphic marker, will allow us to improve the central East Antarctic chronology once more detailed dust records become available using cores from West Antarctica and from relatively high accumulation sites in East Antarctica.

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