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Direct North-South synchronization of abrupt climate change record in ice cores using beryllium 10

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Abstract

A new, decadally resolved record of the ^{10}Be peak at 41 kyr from the EPICA Dome C ice core (Antarctica) is used to match it with the same peak in the GRIP ice core (Greenland). This permits a direct synchronisation of the climatic variations around 41 kyr BP, independent of uncertainties related to the ice age-gas age difference in ice cores. Dansgaard-Oeschger event 10 is in the period of best synchronisation and is found to be coeval with an Antarctic temperature maximum. Simulations using a thermal bipolar seesaw model agree reasonably well with the observed relative climate chronology in these two cores. They also reproduce three Antarctic warming events between A1 and A2.

1 Introduction

Precise correlation of northern and southern hemisphere ice core records is of critical importance for the understanding of the dynamics of abrupt climatic changes during the last glacial period that have been inferred from the isotopic records in the GRIP (Johnsen et al., 1992; Dansgaard et al., 1993) and GISP2 (Grootes et al., 1993) ice cores from Central Greenland during the late glacial period. Twenty-five of these so-called Dansgaard/Oeschger events (DO events), characterized by a rapid warming and a more gradual return to glacial conditions, were identified between 110 and 14 kyr BP (1 kyr BP is 1000 years before present) (North GRIP members, 2004). That the largest of these abrupt changes seemed to have attenuated counterparts in Antarctica was noted from a visual comparison with the isotopic record from the East Antarctic Vostok core (Jouzel et al., 1994). Based on the comparison of the δ^{18}O records in air bubbles, Bender et al. (1994) proposed a first synchronization of Greenland and Antarctic large glacial events for the period >46 kyr BP. Uncertainties of about ±3 kyr in relative dating, due in large part to the ice age – gas age uncertainty, made it impossible to determine whether events in Greenland and Antarctic ice cores were in phase or out of phase

(Bender et al., 1994), or in a lead/lag relationship (Steig and Alley, 2002; Schmittner et al., 2003).

Yiou et al. (1997) proposed a direct synchronization based on the use of the ^{10}Be peak measured in 5 ice cores around 40 kyr BP. The advantage of this method is that the parameter is recorded in the ice rather than in the gases, and that it is independent of climate. In Greenland ice this ^{10}Be peak straddles DO 10 (Yiou et al., 1997) which occurred between the two major DO events 8 and DO 12. Similar to the Greenland isotopic record, there are 3 secondary isotopic maxima between the Antarctic counterparts of DO 8 and DO 12 (Yiou et al., 1997), later named A1 and A2 (Blunier et al., 1998), with the ^{10}Be peak occurring slightly after the second of these events. This would imply that the event corresponding to DO 10 occurred in Antarctica slightly earlier than in Greenland. However, the precision of the timing was limited by the fact that the ^{10}Be and δD measurements were made on two different Vostok cores (see note added in proof of Yiou et al., 1997).

Since then the synchronization of Greenland and Antarctic glacial records has focused on the use of gas indicators. Rapid methane concentration changes proved very powerful for this purpose. Taking advantage of the relatively high accumulation at the Byrd site in West Antarctica, which results in low gas age – ice age uncertainty, Blunier et al. (1998) showed that, comparing their starting points, the long lasting DO8 and DO12 warming events lag their A1 and A2 Antarctic counterparts by 2 to 3 kyr. This conclusion was extended to the seven major millennial-scale Antarctic warmings, A1 to A7, clearly identified back to 90 kyr (Blunier and Brook, 2001) and further to A8/DO 23 (Caillon et al., 2003) and A9/DO 24 (Landais et al., 2006). A further, gas-based synchronization was derived from a very detailed comparison of the Vostok and GISP2 δ^{18}O of air records (Bender et al., 1999). It showed that all 15 long and short interstadial events identified in the GISP2 ice core between 38 and 72 ka have counterparts in the Vostok record. On average, all these events are in phase between Greenland and Antarctica (within ±1.3 kyr). Importantly, the contradiction between the methane and δ^{18}O of air approaches appears to be largely a matter of definition simply due to the different shapes of Greenland and Antarctic events. As noted in Blunier and Brook (2001), a comparison of peak temperatures as done in Bender et al. (1999), rather than the leading edges, would also lead to the conclusion that Byrd and Greenland (GRIP and GISP2) events are in phase.

There are two limitations to the use of gases for synchronization: (i) it is applicable only at times where there is a rapid change in the atmospheric value, (ii) the signal is retained at the "close-off" depth (typically 80 m) where the bubbles in the ice are isolated from the overlying atmosphere. This signal is thus both smoothed and shifted compared to records in the ice itself. The resulting age offset is particularly important in low accumulation rate regions, such as the Antarctic plateau, where it can be as large as 5 kyr or more during glacial periods.

Here we show that the cosmogenic isotope ^{10}Be is a parameter that permits the synchronisation of paleoclimatic records, because large changes, caused by the modulation of this intensity by the geomagnetic field and the electric and magnetic fields associated with the out-flowing solar wind, are globally synchronous and independent of climate. The successful application of such a procedure requires that one can extract the ^{10}Be production rate variations from those caused by changes in deposition patterns (which themselves can be dependent on climate), and that the variations are characteristic enough to identify in the records being examined.

The strongest and most reliably identified excursion of high ^{10}Be concentrations is found in several ice cores from Antarctica and Greenland (Yiou et al., 1997; Raisbeck et al., 1987; Beer et al., 1992), as well as in marine sediment cores (McHargue et al., 1995; Castagnoli et al., 1995; Robinson et al., 1995) ~40 000 years ago. We report here new high resolution measurements of this ^{10}Be peak on samples from the EPICA Dome C ice core and combine these with a ^{10}Be record from Greenland in order to determine the relative phasing of temperature changes recorded in Greenland and Antarctica.

2 Experimental procedure

The ice available for this study was in the form of a continuous series of 55 cm strips, each weighing ~300 g. In order to maximize the time resolution, each bag was divided into 5 samples of 11 cm in length, each representing ~9 years The amount of ice per sample (~50 g) is almost an order of magnitude smaller than we have used for previous studies (Raisbeck et al., 1987; Yiou et al., 1997). We therefore have modified somewhat our previous procedure. The samples were melted in 250 ml centrifuge cones, in the presence of 0.25 mg of ^{9}Be carrier. The $Be(OH)_2$ was then precipitated directly with NH_4OH. The precipitate was then washed with water (pH=7), dissolved in a few drops of nitric acid, and transferred to a quartz crucible. The crucible was taken to dryness on a hotplate, then heated to 900° over an electric furnace in order to transform the precipitate to BeO. The BeO was mixed with Nb powder (325 mesh) in the ratio 3:1 and pressed into a 1 mm diameter, 1 mm deep hole in a Mo cathode. This cathode was heated in argon to 1200° for 1 min, then 1800° for an additional minute, in a resistively heated carbon furnace. This procedure was found to give better and more stable ^{9}Be currents in the ion source, as well as reduce ^{10}B interference. The ^{10}Be/^{9}Be ratios were measured at the Gif-sur-Yvette Tandetron based AMS facility, relative to NIST standard SRM 4325, using the certified ratio of 2.68×10^{-11}. For almost all samples, at least 200Be events were recorded, leading to a 1 sigma statistical uncertainty of <7%. Combined with a conservatively estimated 5% machine uncertainty, this leads to an overall uncertainty of ~8.5%.

3 Results

The ^{10}Be peak is centred at 740 m (Fig. 1), and shows structure on centennial and even decadal time scales (Raisbeck et al., 2002). This suggests that the peak is due to a combination of low geomagnetic field intensity and periods of low solar activity. The effects of the latter are enhanced because they occur during the interval of low geomagnetic field intensity (With a "normal" geomagnetic field, a large fraction of cosmic rays with energies <1 GeV are already excluded from the atmosphere, and are thus insensitive to solar modulation). ^{10}Be reaches the Antarctic plateau primarily by dry deposition, and hence concentrations are inversely correlated with the snow accumulation rate. A more appropriate parameter to use when studying production variations is the ^{10}Be flux which is the product of the measured concentrations and the estimated accumulation rates .(Raisbeck et al., 1992).

The δD profile in the interval between 700 and 800 m has 3 features that, based on their position and pattern, are identified as subdued analogues of DO events 9 to 11 observed in Greenland. The ^{10}Be peak straddles the middle one of these events (Raisbeck et al., 2002), exactly as seen in the GRIP core (Yiou et al., 1997). This implies that DO 10 and its Antarctic counterpart are synchronous in the two ice cores.

Figure 2 shows the best match between the ^{10}Be flux curves of Dome C and GRIP. At this depth, the time resolution of the GRIP ^{10}Be measurements is from ∼30–50 years. In order to have comparable resolution, we have therefore resampled both data at 10 year intervals, and smoothed them by summing the first 5 components of a singular spectral analysis (Paillard et al., 1996). Using these smoothed curves for alignment, DO 10 and its counterpart, AIM 10 (AIM = Antarctic Isotope Maximum; EPICA Community Members, 2006) are synchronous in the two records, within our ability to resolve, which we presently estimate as ∼200 years.

This synchronism for DO 10 and its Antarctic counterpart complements and reinforces results based on the methane records of Antarctic and Greenland ice cores (Blunier et al., 1998) which demonstrated that the start of the warming associated with events A1 and A2 lead the warming of DO 8 and 12, respectively by about 1500 years. Because of the duration of A1 and A2, however, this is equivalent to the alternative statement that the maximum temperature during A1 and A2 coincided with that of DO 8 and DO 12, respectively. For the period in between, with the three intervening DO events 9 to 11, a statement regarding the phase relationship between Greenland and Antarctica for other DO events was not possible owing to the noisier δ^{18}O record of the Byrd ice core and the general uncertainty in synchronising shorter events based on gas records.

4 Implications for bipolar seasaw model

The fact that between A1 and A2 three additional Antarctic temperature maxima can be identified in the EPICA Dome C ice core, and that the middle of these can be precisely synchronised with DO 10 in the Greenland ice cores via the ^{10}Be peak, lends strong support for the thermal bipolar seesaw operating throughout the sequence of DO events (Broecker, 1998; Stocker, 1998). This model assumes that temperature anomalies associated with the abrupt climate changes in Greenland occur in antiphase with those in the South Atlantic (Stocker, 1998) due to the effect of the meridional overturning circulation in the Atlantic Ocean on the oceanic meridional heat flux (Crowley, 1992; Knutti et al., 2004). An important extension of the original concept (Crowley, 1992) is that the Southern Ocean acts as a heat reservoir for temperature anomalies and hence, only a damped and time-integrated signal of the abrupt changes in the north is transmitted to Antarctica. The thermal bipolar seesaw involves a typical heat exchange time scale τ. The best correlation between the simulated southern signal based on the seesaw model and the reconstructed Antarctic temperature was obtained with $\tau \approx 1200$ years.

We now test the thermal bipolar seesaw concept with the new synchronisation based on ^{10}Be. The new temperature reconstructions based on thermal fractionation measured on air enclosed in samples from the NorthGRIP ice core (Huber et al., 2006) is used as input to the seesaw model with $\tau=1200$ and 2200 years. In both cases the southern temperature response of the seesaw model in the time period from 44 to 39 kyr BP exhibits all warmings and coolings which are inferred from the δD record from EPICA Dome C (Fig. 2). It is remarkable that not only the general evolution of the simulated signal is in good agreement with variations, but the differing amplitudes of the three Antarctic events are quite well reproduced. This is particularly the case for τ

= 2200 years

For AIM 8–10, the time difference between the predicted and observed temperature profiles is <200 years, which is about the limit of the estimated uncertainty in our synchronization. For AIM 11 the difference is larger. However, the features of the ^{10}Be fluxes around DO 11 und AIM 11 are not unambiguous enough to warrant additional synchronisation points. Such points might well move AIM 11 earlier by several hundred years.

5 Perspectives

It would of course be desirable to apply the ^{10}Be technique to other time intervals of the climate records. Unfortunately, we have not yet identified other events during the last climate cycle which are as dramatic as the ^{10}Be peaks around 40 kyr BP. However, if continuous high resolution ^{10}Be profiles were available, it still should be possible to make such a correlation. For example, at Vostok, it was possible to correlate the centennial scale variations produced by solar modulation with similar variations observed in tree ring ^{14}C (Raisbeck et al., 1998). In principle, therefore, it should be feasible to correlate such variations in Antarctic and Greenland ice. If one accepts that the decadal variations seen in the ^{10}Be profile of Fig. 1 are caused by production, it should also be possible to improve even further the precision of such correlations. As mentioned above, the degree of correlation in Fig. 2 above is limited by the time resolution of the GRIP ^{10}Be record. Thus, an improved Greenland ^{10}Be record should allow an improved correlation. We have proposed, for example, to measure such a profile in the North GRIP core at a resolution comparable to that measured at Dome C. This would thus potentially permit a correlation on a time scale of better than 20 years. As noted earlier, the largest and longest of the ^{10}Be production variations should also be identifiable in marine sediments, further permitting a direct correlation of the ice core and ocean climatic records.

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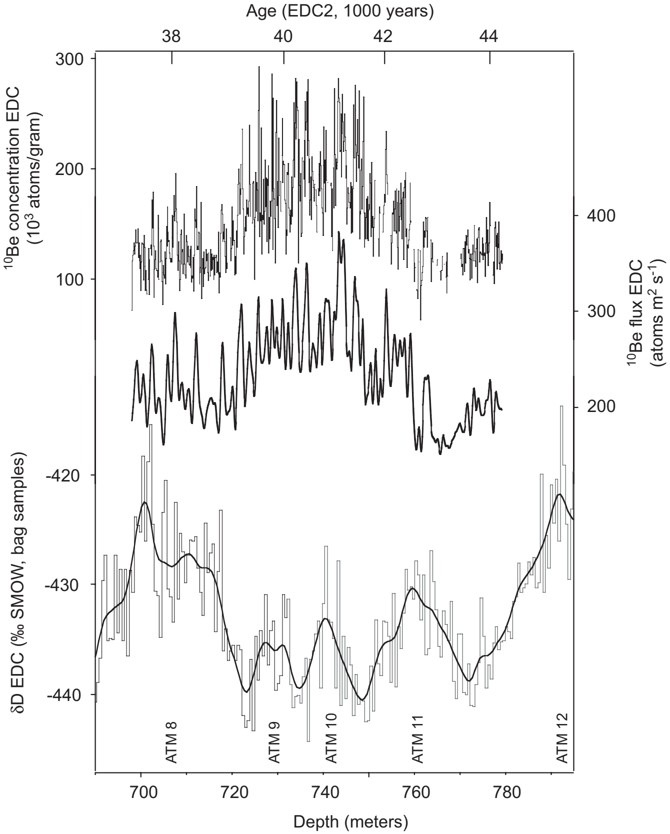

Fig. 1. ^{10}Be concentrations, ^{10}Be flux and deuterium ratio and its running average as a function of depth and age (EDC2) in the EPICA Dome C ice core. The ^{10}Be flux has been calculated using accumulation rates (EPICA Community Members, 2004), and smoothed by summing the 5 first components provided by the singular spectral analysis of the record (Paillard et al., 1996). Antarctic equivalents of DO events 8 to 12 are labeled AIM 8 to 12, where AIM stands for Antarctic Isotope Maximum (AIM).

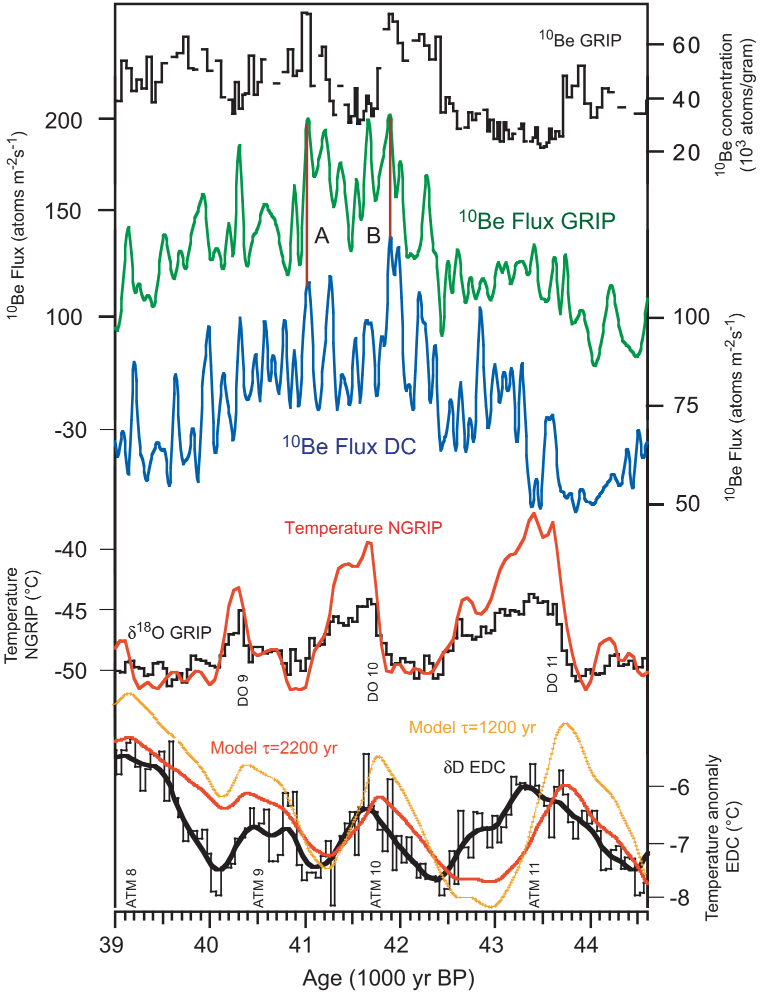
^{10}Be GRIP
60
40
20
^{10}Be concentration (10^3 atoms/gram)
200
150
100
^{10}Be Flux (atoms m^{-2}s^{-1})
A
B
^{10}Be Flux GRIP
100
75
50
^{10}Be Flux (atoms m^{-2}s^{-1})
^{10}Be Flux DC
-30
-40
-45
-50
Temperature NGRIP (°C)
Temperature NGRIP
δ^{18}O GRIP
DO 9
DO 10
DO 11
Model τ=1200 yr
Model τ=2200 yr
δD EDC
-6
-7
-8
Temperature anomaly EDC (°C)
ATM 8
ATM 9
ATM 10
ATM 11
39
40
41
42
43
44
Age (1000 yr BP)

Fig. 2.

Fig. 2. Synchronization of the GRIP and EPICA Dome C records on the SS09sea time-scale (North GRIP Members, 2004), using the two ^{10}Be tie points labeled A and B to align EDC2 with SS09sea, which corresponds to a shift of 815 years. ^{10}Be concentrations are mainly from Yiou et al. (1999) with additional measurements from Wagner et al. (2001) and Muscheler et al. (2004). Corrections for ^{10}Be retained on filters for GRIP samples processed with 0.45 micron filters have been made as described in Yiou et al. (1999). GRIP fluxes, calculated using the accumulation estimates of Johnsen et al. (2001) after applying a correction for the change in the isotopic composition of the ocean (Waelbroeck et al., 2002), are smoothed by singular spectral analysis as in Fig. 1. The two following curves show δ^{18}O at GRIP and estimated temperature change as given by Huber et al. (2006) at NorthGRIP, these two records being placed on a common timescale (North GRIP Members, 2004) and scaled in such a way that the δ^{18}O record corresponds to the temperature change derived using the present-day temperature isotope relationship as observed along traverses (Johnsen et al., 1989). The bottom curves represent the Dome C temperature change (Stenni et al., 2004) without source correction and its running average. This can be compared with the predicted southern temperature response using the thermal bipolar seasaw model (Stocker and Johnsen, 2003), with a τ of 1200 and 2200 years.