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Abstract. Concentrations of the cosmogenic isotope ^{10}Be have been measured in more than 1350 samples from the Greenland Ice Core Project (GRIP) ice core drilled at Summit, Greenland. Although a dust-associated component of ^{10}Be retained by 0.45 μm filters in some of the samples complicates the interpretations, the results confirm that the first-order origin of ^{10}Be concentration variations is changes in precipitation rate associated with different climate regimes. This effect is seen not only between glacial and interglacial periods, but also during the shorter "Dansgaard-Oeschger" interstadials. By contrast, the ^{10}Be data do not support the interpretation of rapidly varying accumulation (i.e., climate) during the last interglacial. They can, however, be used to help place limits on the origin of the ice in these events. After taking into account variable snow accumulation effects, variations in the ^{10}Be flux are observed, probably caused by solar and geomagnetic modulation, but possibly also by primary cosmic ray variations. The most dramatic is a ^{10}Be peak ~40,000 years ago, similar to that found in the Vostok ice core, thus permitting a very precise correlation between climate records from Arctic and Antarctic ice cores. The ^{36}Cl/^{10}Be ratio (considering either "total" or only ice-associated ^{10}Be) shows significant variability over the whole core depth, thus confirming the difficulty in using this parameter for "dating" ice cores.

Introduction

Cosmogenic isotopes such as ^{10}Be and ^{36}Cl in polar ice cores can potentially give information on the history of cosmogenic production rates in the past [*Raisbeck and Yiou*, 1984] and thus on the three parameters which control this production, namely (1) primary cosmic ray intensity, (2) solar modulation, and (3) geomagnetic modulation. In addition, it has been proposed that ^{10}Be can help establish chronologies for such ice cores by (1) permitting estimates of paleoaccumulation rates where annual layers are not otherwise available, either because of low initial precipitation rates or due to thinning of ice near bedrock [*Yiou et al.*, 1985] and (2) providing distinctive features which can be used as stratigraphic markers in the ice [*Raisbeck et al.*, 1987a]. In the framework of the Greenland Ice Core Project (GRIP), a continuous (except during the last glacial-interglacial transition) series of samples were allocated for cosmogenic isotope (^{10}Be, ^{36}Cl, and ^{26}Al) analyses. We report here on the present status of these measurements.

Experimental Procedures and Problems

The samples, consisting of 55 cm sections of ice weighing ~1 kg in the Holocene, or ~0.5 kg in the pre-Holocene, were either treated immediately in the field or reserved as ice for subsequent processing in the laboratory. For logistical reasons imposed by the way the ^{36}Cl sampling was chosen, essentially all the laboratory processing to date has been carried out at Zurich. In both cases the ice was melted in the presence of 0.3 mg of ^{9}Be, 2 mg of Cl, and 0.01 mg of ^{27}Al carriers.

The meltwater was first acidified (pH = 2), filtered (see below), and passed through a cation ion exchange resin (AG 50W-X8 (H^+)) to retain the Be and Al, in a procedure similar to that used earlier [*Raisbeck et al.*, 1981], and then through an anion exchange column for extraction of Cl, as described elsewhere [*Baumgartner et al.*, this issue]. Alternate cation resins were reserved for treatment at Orsay and Zurich, respectively. Be was extracted from the cation columns using 1 N HCl and the resins (containing the Al) reserved for later analyses of ^{26}Al. The Be was transformed into BeO and ^{10}Be/^{9}Be analysis of alternate samples carried out at the accelerator mass spectrometry (AMS) facilities at Gif-sur-Yvette [*Raisbeck et al.*, 1987b] and Zurich [*Suter et al.*, 1989], both relative to the National Institute of Standards and Technology (NIST) standard SRM 4325. At Gif-sur-Yvette, ^{10}Be/^{9}Be backgrounds are $<2 \times 10^{-15}$ [*Raisbeck et al.*, 1994], which is negligible compared to the ratios being measured. In most cases, at least 400 ^{10}Be ions were counted, giving a statistical uncertainty of <5%. This, together with a conservative estimate of 5% for machine variability [*Raisbeck et al.*, 1987b], results in measurement uncertainties of ~7%. The uncertainties of the Zurich measurements are a combination of counting statistics, the measured

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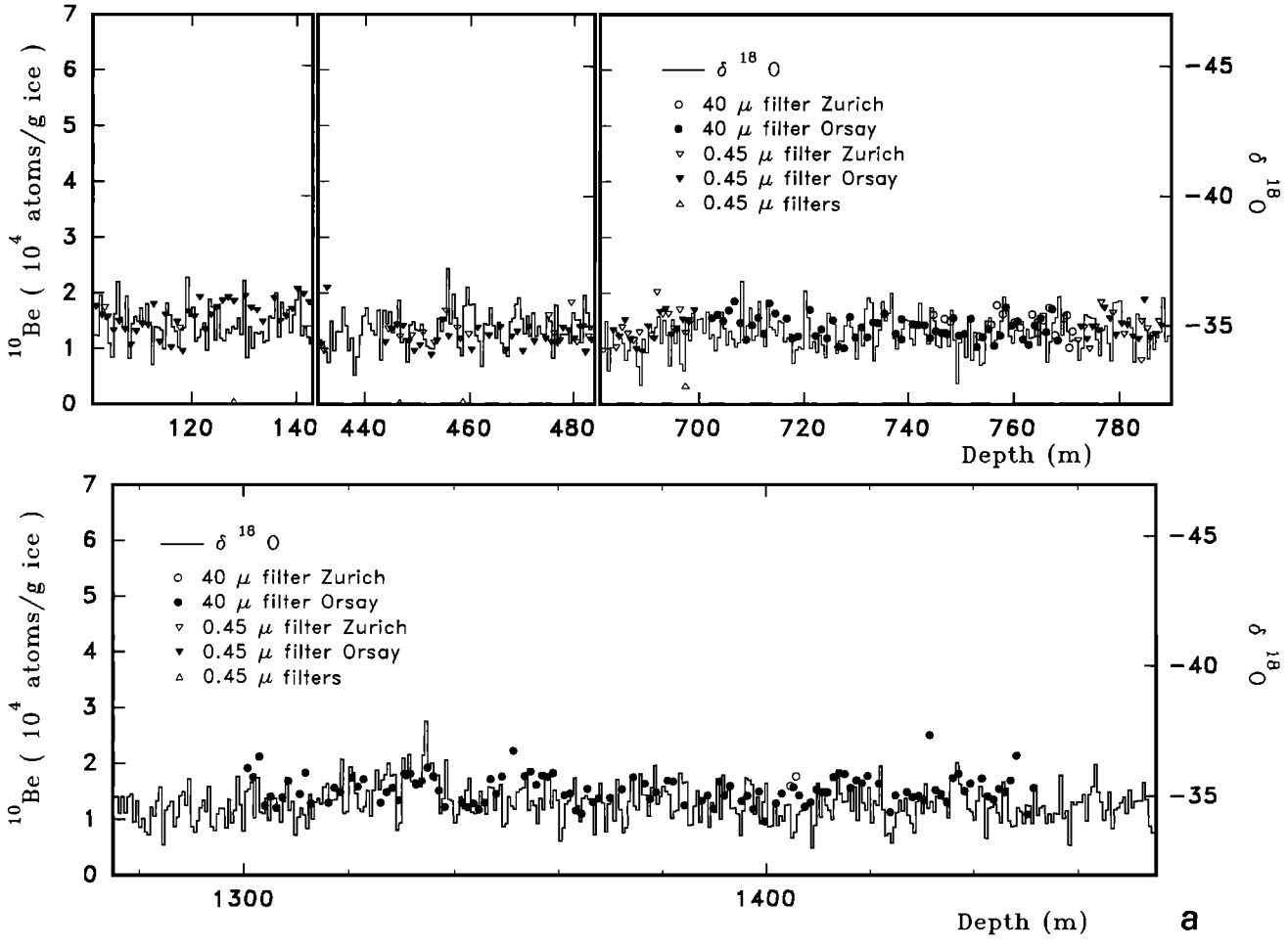

Figure 1. Beryllium 10 concentrations in samples studied from the GRIP core. Circles represent samples processed with 30 or 45 μm filters; inverted triangles represent samples processed with 0.45 μm filters. Solid symbols are samples measured at Gif-sur-Yvette; open symbols are samples measured at Zurich. Upright open triangles represent ^{10}Be leached from the 0.45 μm filters, expressed as a concentration of the quantity of ice filtered. The histogram is the δ^{18}O record [*Dansgaard et al.*, 1993]. Data shown are from (a) Holocene ice, (b and c) last glacial period (LGP) ice, and (d) pre-LGP ice.

reproducibility of several measurements of the same sample and of standards (standards are measured after every five samples), and where necessary, a background correction, the latter of which is discussed further below.

To date, a total of 1355 samples (1025 at Gif and 330 at Zurich) have been analyzed for ^{10}Be (Figure 1). Although preliminary results, including observation of a ^{10}Be "peak" at ~2240 m, have been presented orally since 1993, formal publication has been delayed because of two unforeseen complications. First, a detailed comparison of the early results revealed that the measurements at Zurich were systematically higher than those made at Gif-sur-Yvette. Detailed reanalysis of the Zurich data showed that the background correction had been underestimated for samples with high boron content. This problem concerns 134 of 359 samples measured in Zurich. A new procedure was developed for the data in question. Measurements with corrections >20% (29 out of 134) were excluded from further analysis. The median correction for the remaining samples was 9%. The formal mean (median) final sample error distribution, including these corrections, is 8% (7%).

The number of results corrected in this way represents only ~8% of the total data set. However, because those samples included several from the region of the observed ^{10}Be peak, and demonstrating the existence and exact form of this peak was considered very important, it was agreed that the French group would resample the core at six critical levels in the peak region, process the samples at Orsay, and measure them at Gif-sur-Yvette. The results for these duplicate samples are shown in Table 1. At the time, there appeared to be quite satisfactory agreement. Ironically, this good agreement may now raise additional questions, because of the second problem, discussed below.

Shortly after the aforementioned background problem had been resolved, the Orsay group discovered that measurements on samples from near the bottom of the core were giving systematically different results depending on whether the samples had been processed in the field or at Zurich. In accord with the mutually agreed upon protocol, the samples in the field were filtered through 45 μm filters, similar to those used previously at Orsay to isolate micrometeorites [*Yiou and Raisbeck*, 1987]. In contrast, and unknown to the Orsay authors, the samples treated at Zurich were filtered through 0.45 μm Millipore filters. Subsequent tests by the Swiss group showed that

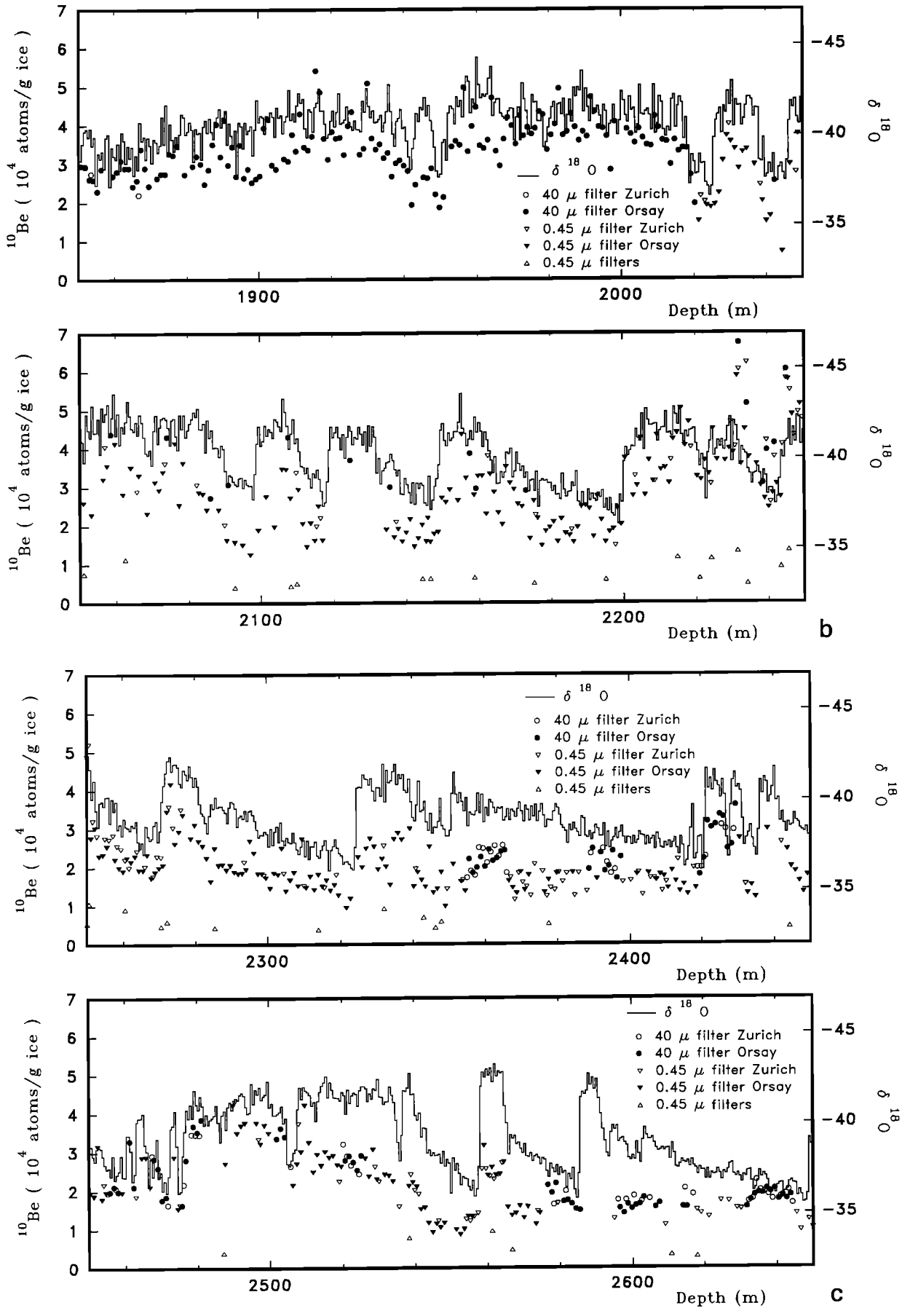

Figure 1. (continued)

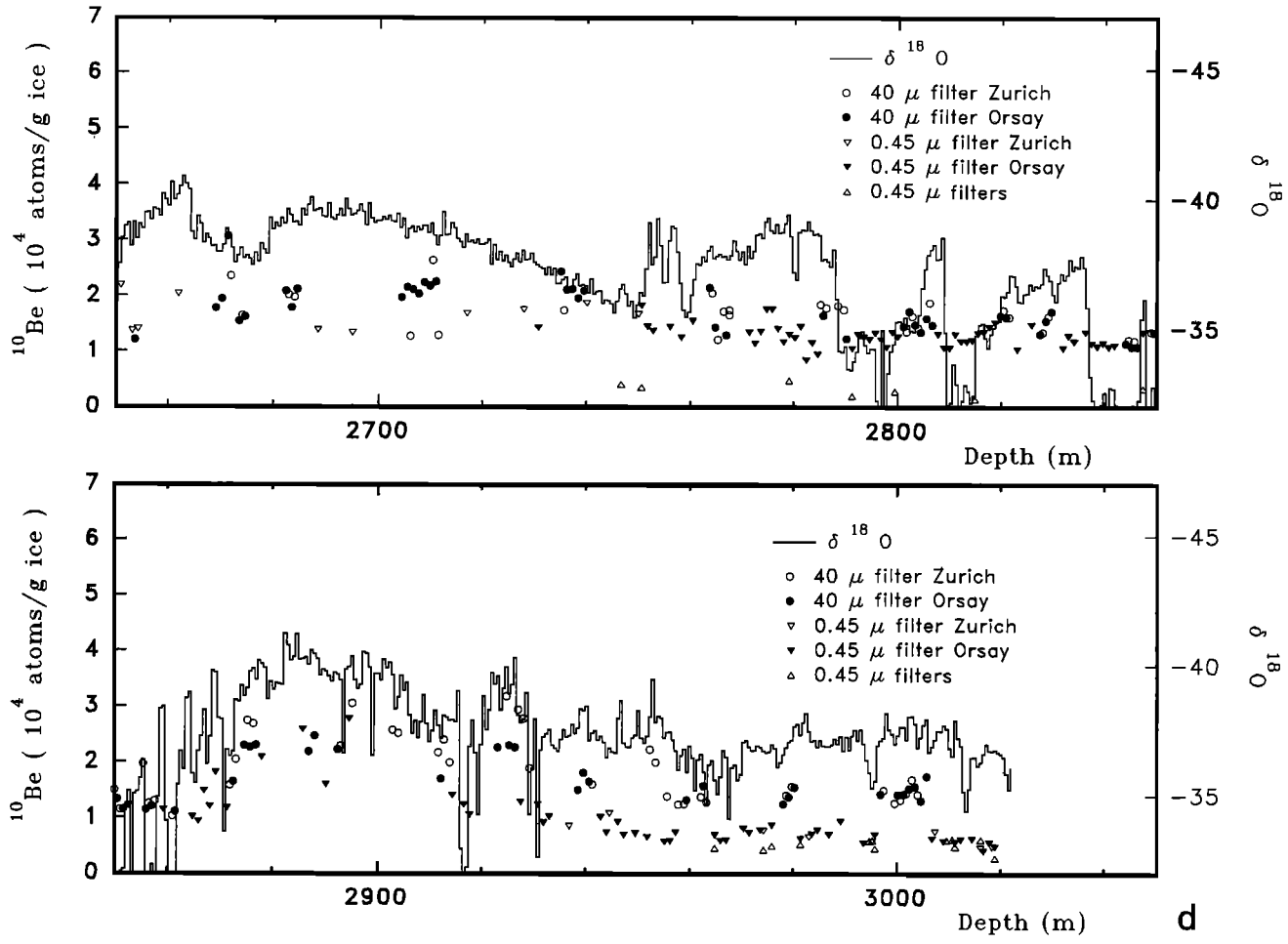

Figure 1. (continued)

a significant quantity of ^{10}Be (but not ^{9}Be carrier) could be leached off the 0.45 μm filters with 1 *N* HCl [*Baumgartner et al.*, 1997]. It was observed that the sum of ^{10}Be from the meltwater plus filters, for samples processed at Zurich, was similar to the meltwater ^{10}Be in adjacent samples processed in the field (Figure 1). It is thus hypothesized that for samples treated in the field, the "dust" in the ice was retained on the ion exchange column, and the ^{10}Be leached off this dust by the 1 *N* HCl used to extract Be from the resin. Since no significant ^{9}Be carrier was leached off the filters, it seems reasonable to conclude that the ^{10}Be associated with the filtered particles was either already attached to them when the particles were deposited in the ice or adsorbed on them during the melting process (before equilibrium with ^{9}Be had occurred). Further investigations showed that while the phenomenon was most pronounced for samples from the bottom ~100 m of the core, the particle-associated ^{10}Be was significant (15–30%) over all of the glacial period, while apparently less important (<5%) for Holocene samples [*Baumgartner et al.*, 1997].

We thus find ourselves in the awkward position of having a

Table 1. Results of Duplicated Samples

Bag Number	Samples Processed at Zurich With 0.45 μm Filters			Samples Processed at Orsay With 30 μm Filters: Meltwater	Orsay/Zurich Ratio
	Meltwater*	Filter†	Total		
4058	6.02 ± 0.36	···†	···	6.72 ± 0.40	···
4062	6.20 ± 0.40	0.48 ± 0.03	6.68 ± 0.43	5.13 ± 0.31	0.77 ± 0.07
4070	···‡	0.24 ± 0.08	···	3.08 ± 0.18	···
4072	4.17 ± 0.24	0.64 ± 0.06	4.81 ± 0.24	3.93 ± 0.24	0.82 ± 0.06
4076	3.75 ± 0.19	0.28 ± 0.02	4.03 ± 0.21	4.11 ± 0.25	1.02 ± 0.08
4082	5.78 ± 0.30	1.36 ± 0.08	7.14 ± 0.38	6.02 ± 0.36	0.84 ± 0.07

Values are in units of 10^4 atoms ^{10}Be per gram of ice.
*Value after correction for ^{10}Be-induced background.
†Filter not found.
‡No value given because background correction greater than 20%.

data set made up of two types of measurements. On one hand, we have results for samples processed in the field, which apparently includes both ice-associated and particle-associated ^{10}Be. On the other hand, we have, for samples processed at Zurich, results for ^{10}Be associated with the ice only plus, in 40 cases, the ^{10}Be associated with the filtered particles. We also have 19 samples where we presently have results only for the filters. Unfortunately, even the significance of the separate components in these latter cases is unclear, because we cannot exclude the possibility that a fraction (perhaps variable) of the particle-associated ^{10}Be was leached off in the meltwater before passing through the ion exchange columns.

As alluded to earlier, the situation is complicated even further by the results for the duplicate samples given in Table 1. For different reasons, a direct comparison can be made for only four of these samples. However, in three out of these four, the Orsay measurements, made on samples processed with 30 μm filters (and thus presumably representing "total" ^{10}Be), are significantly smaller than the "total" (meltwater plus filter) of the Zurich measurements, made on samples processed with 0.45 μm filters. Assuming that this is not due to underestimated experimental errors, it may indicate that the sum of ^{10}Be in meltwater plus 0.45 μm filters is not exactly the same as in meltwater processed with coarse (30 or 45 μm) filters. Additional experiments will be needed to clarify the situation. Nevertheless, even in its present somewhat unsatisfactory state, we believe the available data set contains a significant amount of useful information as described below.

Results and Discussion

The basic data are shown in Figure 1. For convenience the discussion is separated into three parts, corresponding to (1) Holocene ice (Figure 1a), (2) last glacial period (LGP) ice (Figures 1b and 1c), and (3) pre-LGP ice (Figure 1d). A fourth subsection examines the ^{36}Cl/^{10}Be ratios.

Holocene Ice

As indicated earlier, one of the main objectives of this work is to investigate ^{10}Be production variations caused by solar and geomagnetic modulation effects. To do this effectively will require a continuous ^{10}Be profile. Because of the way the samples were divided, and the limited number of samples measured to date by the Swiss group, we have few sections of continuous results. In the Holocene the only relatively continuous section is from 742–788 m (~1780–2066 B.C.). We performed spectral analyses of the ^{10}Be time series in this interval using three spectral techniques described by *Yiou et al.* [this issue, and references therein], i.e., Monte-Carlo singular spectrum analysis (MC-SSA), the maximum entropy method (MEM), and the multitaper method (MTM). The use of multiple techniques is an important test to verify the robustness of the results. We also varied the characteristic parameters of each of these methods to insure the stability of the results. The data are regularly sampled in depth (δp = 0.55 m) but not in time; thus we interpolated the time series on a regular temporal grid using several techniques [*Yiou et al.*, 1996], with a sampling rate of 3 years, the approximate time interval of each experimental data point.

The spectral results are shown in the three panels of Figure 2. The three techniques (MC-SSA, MEM, and MTM) consistently show periodic components near 19 years and 10.5 years. The 19 year peak seems more stable and more significant (with the MTM F test) than the 10.5 year peak. Peaks at other periodicities appear with MEM and MTM (top and middle plots), but the red noise test of MC-SSA shows that they are within the 90% noise bounds (bottom plot). We stress the fact that the number of data points is relatively small (N = 70). Thus the periodicities are average approximations with 1–2 year uncertainties.

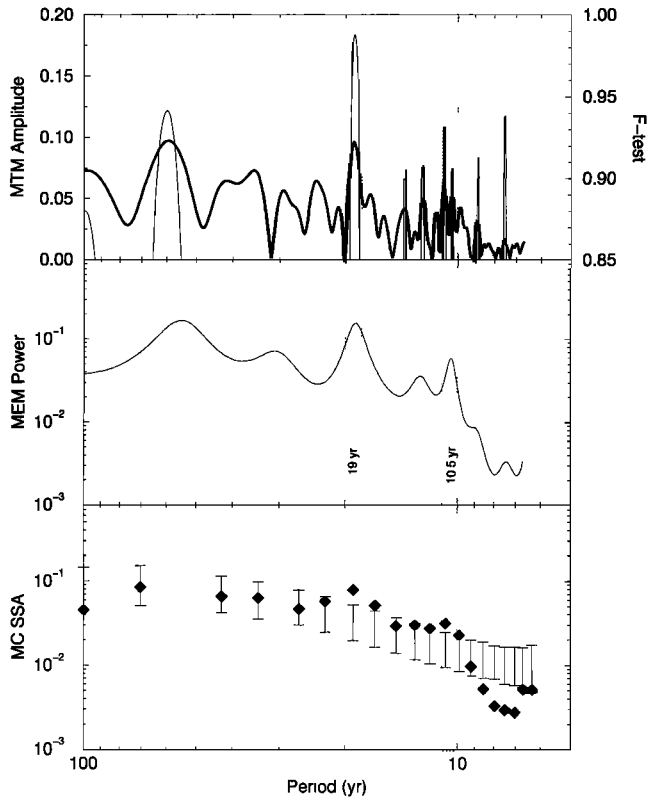

Figure 2. Spectral analyses of a continuous section of the ^{10}Be Holocene record over the depth range from 742 to 788 m (1780–2066 B.C.). The horizontal axis is common to the three panels and shows periods on a logarithmic scale. (Top) Multitaper method (MTM) harmonic analysis with seven tapers and a frequency bandwidth of $N\Omega$ = 4. The thin line represents the amplitude and the thick line is the statistical F test. (Middle) Maximum entropy method (MEM) analysis with an autoregressive order of M = 20. (Bottom) the results of Monte Carlo singular spectrum analysis (MC-SSA) with an embedding dimension M = 20. The vertical bars represent the range of singular spectrum variations for red noise simulations; the diamonds associated with those bars represent components which are 90% likely not to be accounted for by such noise.

The observed periodicities are close to the 11 year sunspot, and 22 year magnetic solar cycles, which give rise to ^{10}Be production variations [*Raisbeck and Yiou*, 1980; *Beer et al.*, 1990]. The 19 year cycle is also close to the 18.6 year luninodal cycle [*Danby*, 1962]. This could possibly be related to climatic effects. However, a similar analysis of δ^{18}O over the same period shows only very weak indications of this periodicity. This reinforces the interpretation of these ^{10}Be cycles as being due to production variations, and gives hope that where sampling resolution permits, it may be possible to observe evidence of these cycles over much of the Holocene part of the GRIP core.

We have also carried out spectral analysis (not shown) of the

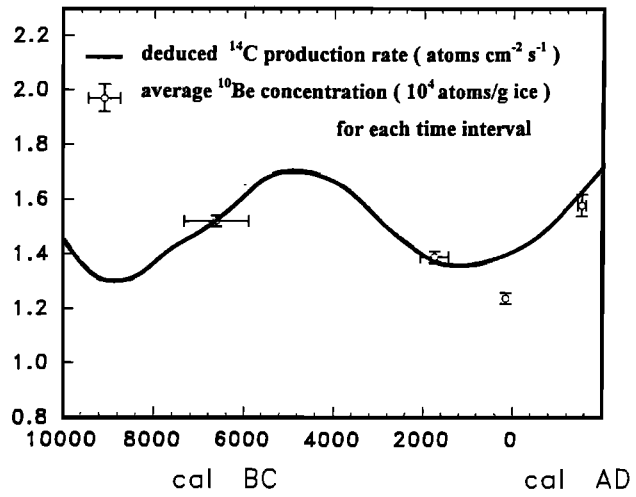

Figure 3. Average ^{10}Be concentrations for the four sequences of Holocene samples studied and shown in Figure 1a. Vertical bars represent one standard deviation of the mean; horizontal bars represent the time range of each series, using the chronology of *Dansgaard et al.* [1993]. The solid line represents the long-term trend of ^{14}C production rate, as estimated by *Stuiver and Braziunas* [1993]. The numerical identity of the ordinate scales is coincidental.

^{10}Be data for the period 1300–1450 m (5890–7340 B.C.), where each sample represents ~5 years, and we have data for every second sample. The only clear periodic component observed was ~52 years. We did not find any evidence for 80 or 200 year cycles, which are frequently observed in analyses of tree ring ^{14}C data [*Damon and Sonett*, 1991].

To investigate longer timescale effects, we have averaged the available data for each of the four time periods for which we have data (Figure 3). The horizontal error bars represent the time period covered, and the vertical error bars represent the standard deviation of the means for each series of measurements. This assumes that the fluctuations around the means are distributed in a Gaussian fashion, which is probably not strictly true. Nevertheless, we believe the observed variations in the means are significant. An additional complication, of course, is that the younger two series of samples, and a small portion of those at ~750 m, were processed with 0.45 μm filters, while the others were processed with 45 μm filters. As mentioned, however, some preliminary results [*Baumgartner et al.*, 1997] suggest that the dust contribution to ^{10}Be in Holocene samples is usually <5%. If this is confirmed, then corrections for the dust contribution should not significantly alter the basic trends in Figure 3.

Variations in the ^{14}C/^{12}C ratio in the Holocene atmosphere, as deduced from tree ring data, have often been attributed to changes in the Earth's geomagnetic field intensity, although climatic effects have also been proposed. While originally attributed to geomagnetic changes in the Holocene itself [*Damon et al.*, 1978], subsequent analyses suggested that much of the slow ^{14}C variation in the Holocene was an ocean-moderated "relaxation" response to increased production during the late glacial period [*Stuiver and Braziunas*, 1993]. Previous ^{10}Be profiles in Arctic ice have, in fact, been cited as evidence that Holocene geomagnetic intensity variations are not sufficient to alone explain the ^{14}C data [*Beer et al.*, 1984a, 1988]. In Figure 3 we thus also show the most recently deduced production variations in the Holocene that would be necessary to account for the observed ^{14}C/^{12}C profiles in tree rings [*Stuiver and Braziunas*, 1993]. There are two problems with such a comparison: (1) while the ^{14}C signal is global, it is not known how sensitive the ^{10}Be deposited at Summit is to changes in the geomagnetic field intensity [*Raisbeck and Yiou*, 1988], and (2) as mentioned above, the ^{14}C system has a long memory and therefore reflects to some extent also the pre-Holocene production history. With the above caveats in mind, however, it appears from Figure 3 that the average ^{10}Be in three of the intervals is consistent with the deduced ^{14}C production curve, while that in the fourth implies a smaller production around 0 A.D. Additional results will certainly be necessary to substantiate this conclusion.

Last Glacial Period (LGP) Ice

One can observe that the ^{10}Be concentrations in the last glacial period (Figures 1b and 1c) are generally 2–3 times larger than the Holocene values (Figure 1a). This phenomenon, which has been noted previously [*Raisbeck et al.*, 1981; *Beer et al.*, 1984b], has generally been attributed to precipitation rate changes between the two periods. While such an explanation is very plausible for the Antarctic plateau, where depositions of aerosols (and thus ^{10}Be) are dominated by dry fallout [*Raisbeck and Yiou*, 1985], it is more difficult to understand for Greenland, where dry fallout is probably minor and moisture transport is often through storms coming from lower latitudes. In regions today where wet fallout of radioactive isotopes dominates, concentration tends, to a first approximation, to be independent of precipitation rate [*Lal and Peters*, 1967]. Thus the ^{10}Be results may suggest that in glacial times, hemispheric or global precipitation was reduced. This would, of course, result in a corresponding increase in ^{10}Be concentration for all precipitation.

One can also observe in Figures 1b and 1c that even in the last glacial period itself, there are significant fluctuations in ^{10}Be concentration, and once again these are generally strongly correlated with periods of less negative δ^{18}O (so-called Dansgaard-Oeschger events). Once again, we attribute these fluctuations to a climatic effect, most probably precipitation rate changes. These fluctuations make it difficult to use ^{10}Be concentration to observe production rate changes. To try to overcome this, we adopt the procedure used previously for the Vostok core [*Raisbeck et al.*, 1992] and plot in Figure 4 the flux of ^{10}Be. This requires an independent estimate of past accumulation rates. We have used two different estimates. The first is based on experimental layer counting down to 2440 m [*Hammer et al.*, 1997] combined with a theoretical "thinning function" [*Dahl-Jensen et al.*, 1993]. The second is based on the semiempirical relationship: acc = 0.23 exp (0.144 (δ^{18}O + 34.8)) [*Dansgaard et al.*, 1993].

For the 32 samples of the LGP where we have data for both ice and filter, the average ratio (^{10}Be in filter)/(^{10}Be in ice) is 0.25 ± 0.09. For samples processed at Zurich where we do not yet have filter measurements, we have thus multiplied the ^{10}Be ice concentrations by 1.25 ± 0.09 to calculate the fluxes of Figure 4. We emphasize that we do not consider this procedure (which increases most of the concentration uncertainties to approximately ±10%) to be a final solution, but simply a first-order attempt to reduce the inhomogeneities in the present data set. For the five levels where we have duplicate measurements (Table 1), we have taken the average of the Orsay and

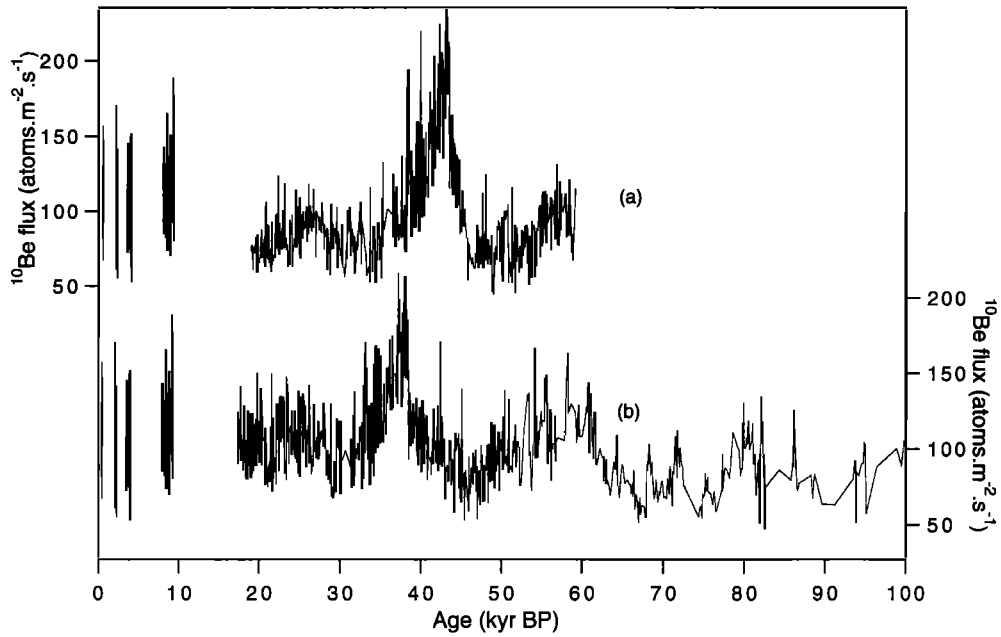

Figure 4. Calculated ^{10}Be flux from measured ^{10}Be concentrations as a function of age. Accumulation rates and chronology are adapted from *Hammer et al.* [1997] (curve a) and *Dansgaard et al.* [1993] (curve b). Samples treated with 0.45 μm filters are the sum of ^{10}Be from meltwater plus filter, or meltwater multiplied by a correction factor (see text).

Zurich values (the latter corrected for both background and dust contributions).

There are two differences between the ^{10}Be fluxes curves of Figure 4. First, using the *Hammer et al.* [1997] derived accumulation rates, the Holocene flux is significantly larger than the average LGP flux. At first glance this would seem to be a weakness of the Hammer et al. chronology. Nevertheless, this is not obvious, since it is not evident that for two periods of such different climate, the ^{10}Be deposition at a given location should remain constant. Second, there is in general more short-term variability using the Hammer et al. derived accumulation rates. This may be due to the fact that the Hammer et al. layer thicknesses are, for the moment, only available for selected climatic intervals, ranging in time from 140 to 3000 years. Thus any short-term accumulation changes which take place during these intervals (as possibly implied by δ^{18}O fluctuations) are not properly taken into account.

Thus, on the basis of the above two criteria, the ^{10}Be data do not favor one chronology over the other. Indeed, one might even argue that the differences between these two chronologies are an indication of the remaining uncertainty of the GRIP timescale.

Most important, from the point of view of the present study, both flux curves of Figure 4 show enhanced ^{10}Be deposition around 2240 m (~38,000 B.P. in the timescale of *Dansgaard et al.* [1993] and ~42,500 B.P. in the timescale of *Hammer et al.* [1997]). Evidence for a 1000–2000 year "peak" in ^{10}Be, ~35 kyr B.P., was originally found in Antarctic ice cores from Vostok and Dome C [*Raisbeck et al.*, 1987a], and subsequently at Byrd [*Beer et al.*, 1992]. Some evidence was also found for an enhancement in ^{10}Be at this time in the Camp Century ice core from Greenland [*Beer et al.*, 1992]. However, because of a relatively modest time resolution and missing portions of the core, the latter identification necessarily remained tentative. Suggested origins of this peak include reduced solar modulation [*Raisbeck et al.*, 1987a], increased primary cosmic ray intensity [*Sonett et al.*, 1987; *Kocharov*, 1990], reduced geomagnetic field intensity [*Mazaud et al.*, 1994], or some combination of these [*Robinson et al.*, 1995]. We show in Figure 5 an expanded plot of the GRIP data for this period, using both the Hammer et al. and Dansgaard et al. timescales, together with the Vostok data on the extended glaciological timescale (EGT) [*Jouzel et al.*, 1993]. For the Dansgaard et al. timescale the Summit and Vostok peaks occur very close in time and have very similar duration and bimodal fine structure. Using the Hammer et al. timescale, the peak is somewhat broader and occurs ~6000 years earlier. Nevertheless, it seems reasonable on the basis of the available evidence to identify the peak at 2240 m in GRIP with that seen in the Antarctic cores. If this interpretation is accepted and the Hammer et al. timescale for GRIP is adopted, it would imply a significant revision of the Vostok timescale.

With this identification, it now becomes possible to correlate the stable isotope (i.e., climatic) records of all five ice cores in which the ^{10}Be peak has been seen. This is done in Figure 6, where we adopt the Hammer et al. GRIP timescale, and the others have been adjusted so as to set the ^{10}Be peak at the same time. The considerable similarity of the climatic records, when calibrated in this way (although the features are strongly damped in Antarctica compared to Greenland) lends support to our interpretation. A similar conclusion for Vostok and GRIP has been arrived at using a visual comparison [*Jouzel et al.*, 1994] or the δ^{18}O in trapped air bubbles as a correlation tool [*Bender et al.*, 1994]. In cases where high-resolution records of ^{10}Be are available, the correlation can be made very precisely (~100 years). This can be important, for example, when considering questions such as possible lead or lag times between climatic records from the northern and southern hemispheres.

To illustrate this, in Figure 7 we show an expanded portion

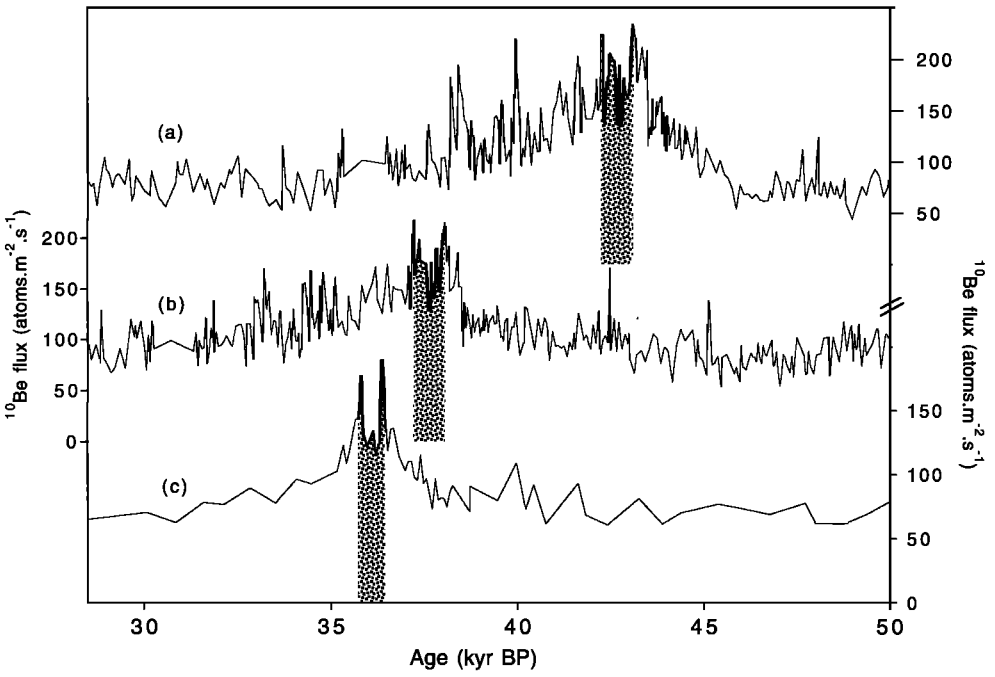

Figure 5. Beryllium 10 flux at GRIP using chronology from *Hammer et al.* [1997] (curve a) and chronology from *Dansgaard et al.* [1993] (curve b), and ^{10}Be flux at Vostok [*Raisbeck et al.*, 1992] using the chronology from *Jouzel et al.* [1993]. Periods of enhanced ^{10}Be flux are shown by shaded areas.

of Figure 6 for the GRIP δ^{18}O and Vostok deuterium records. At GRIP, the ^{10}Be peak straddles the δ^{18}O peak of interstadial 10, which occurs between two major Dansgaard-Oeschger events [*Dansgaard et al.*, 1993]: interstadials 12 (Hengels) and 8 (Denokamp). At Vostok the ^{10}Be peak also occurs between two well-defined deuterium maxima, which on the basis of the above correlation are believed to be counterparts of Greenland interstadials 8 and 12. Interestingly, there are also three secondary deuterium maxima between those two large Vostok events, with the ^{10}Be peak occurring slightly after the second of these events. If this is true, it would indicate that Antarctic climate change slightly leads Greenland climate change at the

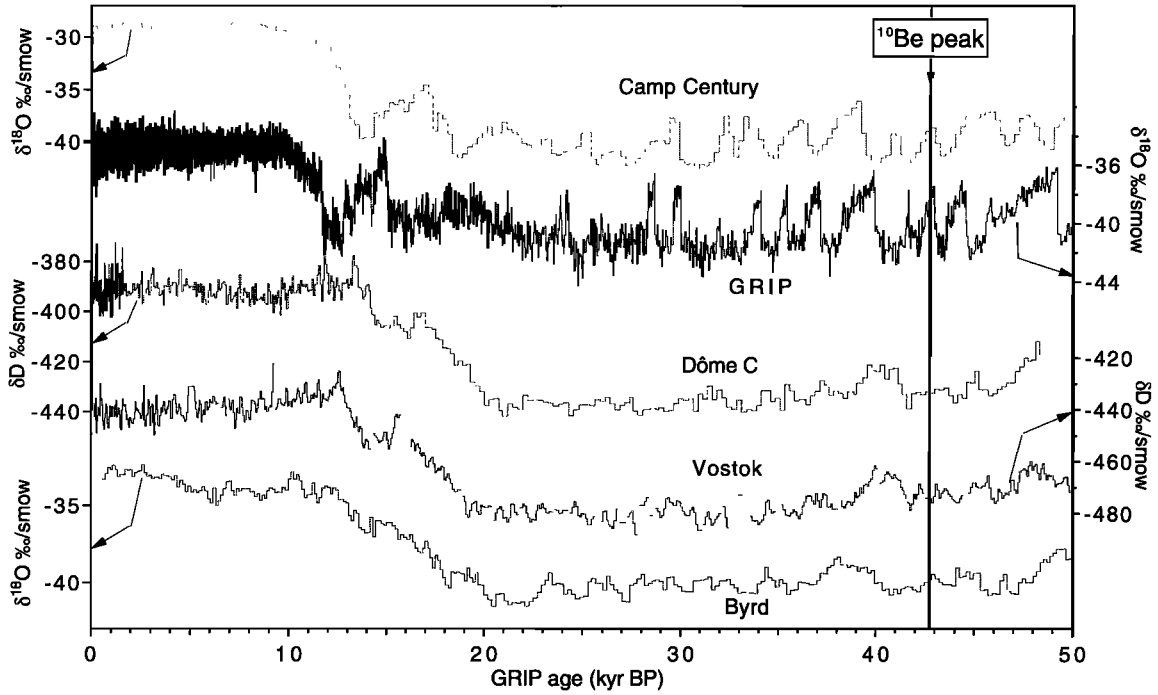

Figure 6. Stable isotope profiles of five polar ice cores in which the ^{10}Be peak at ~40,000 B.P. has been identified. Profiles have been aligned using this peak as a stratigraphic time marker and adopting the GRIP timescale of *Hammer et al.* [1997].

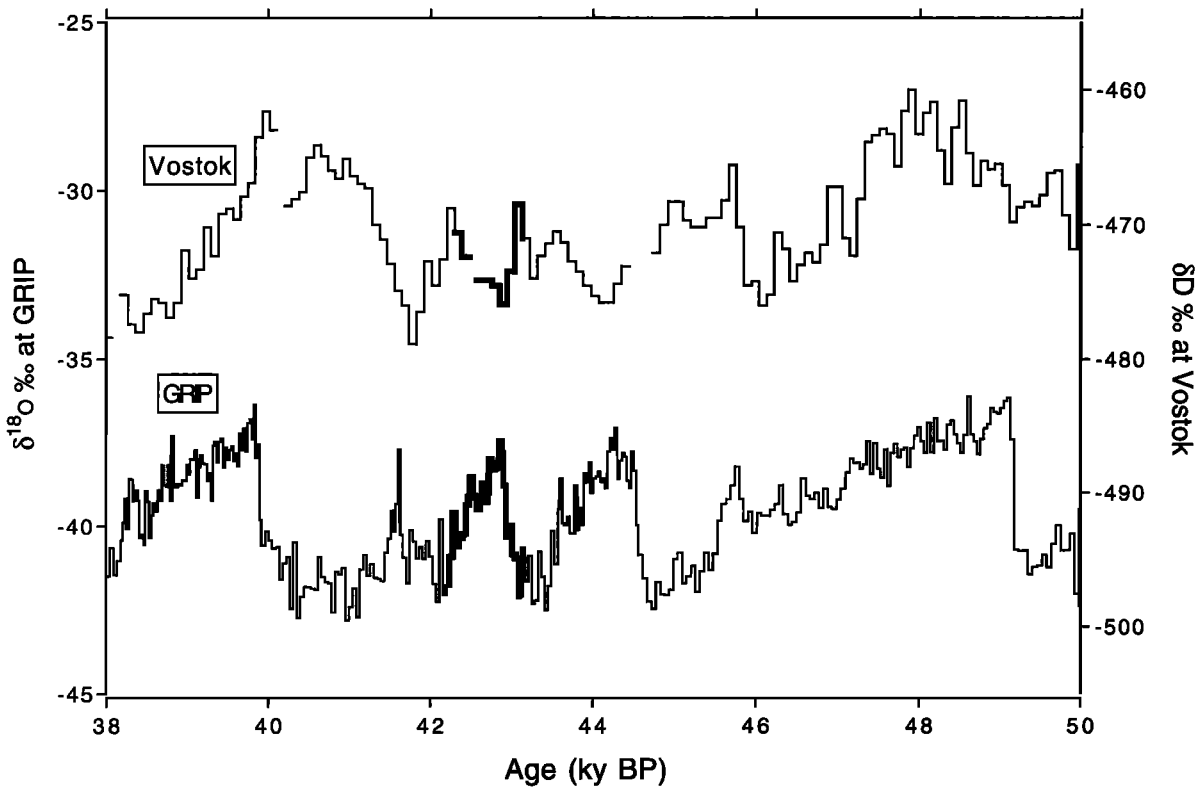

Figure 7. Expanded portion of Figure 6 showing detailed stable isotope profiles for GRIP and Vostok in the region of the ^{10}Be peak. The bold portions of the profiles correspond to the periods of enhanced ^{10}Be flux indicated by shaded areas in Figure 5.

time of the ^{10}Be peak. However, this conclusion must remain tentative at the present time because (1) there are several intervals without deuterium data in the region of the ^{10}Be peak and (2) deuterium has been measured in core 3G while the high-resolution ^{10}Be has been measured in core 4G. These two cores may be offset by a few meters (a few 100 years). We are presently measuring continuous high-resolution deuterium and ^{10}Be profiles in the same core (5G) which should allow us a more definitive conclusion (see note added in proof).

Now that it appears reasonably clear that the ^{10}Be peak at ~40,000 B.P. is global in nature, it will obviously be important to also find evidence of it in marine and lacustrine sediment cores, which would permit one to tie the marine and continental climate records to those of the ice cores. Several possible identifications of this peak in marine cores have indeed been reported recently [*McHargue et al.*, 1995; *Castagnoli et al.*, 1995; *Robinson et al.*, 1995]. The estimated age of the peak in these reports ranges from 32 to 43 kyr B.P., and thus does not yet permit one to choose between the timescales of Hammer et al. and Dansgaard et al. However, if one of the above, or some other, identification of the ^{10}Be peak can be confirmed in a reliably dated reservoir, it would allow ^{10}Be to provide not only a relative but also an absolutely dated stratigraphic horizon for ice cores.

Evidence was also found in the Vostok core for a second ^{10}Be peak around 60 kyr B.P. [*Raisbeck et al.*, 1987a]. A more continuous profile at Vostok (G. M. Raisbeck et al., manuscript in preparation, 1997) shows this peak to also contain structure, though it is smaller in amplitude than the peak at ~40 kyr. While there is no compelling evidence of a ^{10}Be flux peak around 60 kyr in Figure 4, there is some indication of a broad increase around this time. A definitive conclusion will require a more continuous data series and quantitative corrections for the "dust" contributions.

Pre-LGP Ice

While the chronology and continuity of the ice down to ~2700 m (~90 kyr) in the GRIP core seem fairly well established, there is considerable uncertainty before this time. While the region from 2770 to 2865 m was originally thought to represent a climatically variable interglacial period [*Dansgaard et al.*, 1993], this interpretation is now questionable [*Chappellaz et al.*, this issue]. As can be seen in Figure 1d, the available ^{10}Be concentrations in this period are comparable to the Holocene, with no obvious correlation to δ^{18}O. These results thus do not support the interpretation that the more negative δ^{18}O values in this period represent colder (lower accumulation) intervals.

If the "cold" events are incursions of folded ice, then ^{10}Be, like other parameters (gases, chemistry, dust) can be used to place limits on the origin of this ice. For example, there are only a few depths in the GRIP core that have the same combination of δ^{18}O and ^{10}Be as that found in the cold events. One such depth is that corresponding to stage 5d (2750–2790 m), which is also one of the potential depths suggested by the gas data [*Chappellaz et al.*, this issue].

Below 2865 m, the ^{10}Be again increases with more negative δ^{18}O, presumably due to lower accumulation associated with the penultimate glacial period. Below 2900 m, however, there is a considerable decrease in ^{10}Be concentrations for samples processed with 0.45 μm filters. In this region the ^{10}Be associated with dust is almost equal to that associated with the ice. It is clear that any interpretation of the ^{10}Be concentrations (or

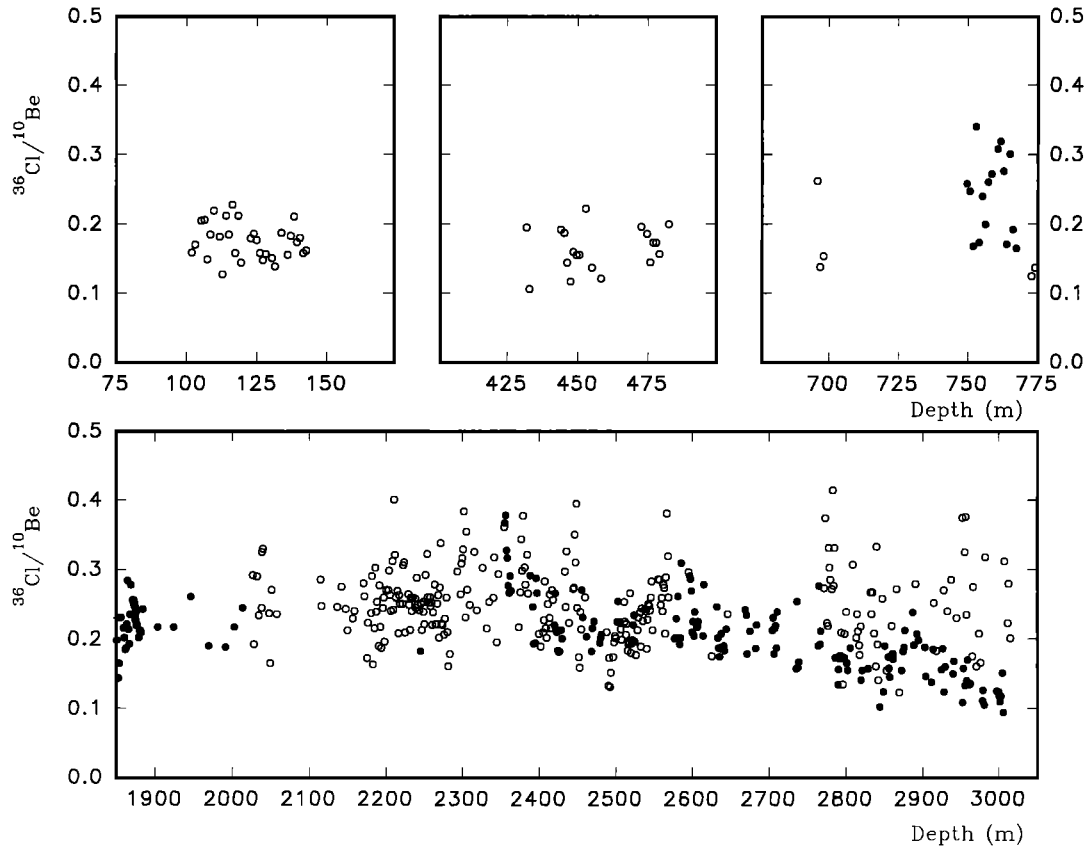

Figure 8. The ratio ^{36}Cl/^{10}Be as a function of depth in core. Solid circles represent samples processed with 30 or 45 μm filters (i.e., "total" ^{10}Be), while open symbols represent only the meltwater component of ^{10}Be for samples processed with 0.45 μm filters.

^{36}Cl/^{10}Be ratios, see below) in this region will depend critically on whether the ^{10}Be associated with the dust should be added to that in the ice. However, we can conclude that the distinctive partition of ^{10}Be between these two phases in the core is quite different than that at any other depth. This could be useful information, for example, when again considering the question of whether last "interglacial" ice having unusual values for other parameters could have come from incursions of folded ice from deeper in the core.

Ratio of ^{36}Cl to ^{10}Be

The available ^{36}Cl results for the GRIP core are presented elsewhere [*Baumgartner et al.*, this issue]. Here we consider the ratio of ^{36}Cl/^{10}Be.

As explained earlier, the ^{36}Cl measurements are made on samples 1.1 m in length, while the ^{10}Be measurements are made on samples of 0.55 m. Thus each ^{36}Cl sample corresponds to two ^{10}Be samples, one to be measured by the Swiss group and the other by the Orsay group. We have at present a relatively limited number of ^{36}Cl results where both corresponding ^{10}Be measurements have been made. For those where we have only a single ^{10}Be measurement, we have, as a first approximation, calculated the ^{36}Cl/^{10}Be ratio assuming the available ^{10}Be concentration is applicable for the entire ^{36}Cl sample. While this will undoubtedly lead to additional scatter in the data, it should not change the basic behavior of this parameter (Figure 8).

At one time it had been hoped that the ^{36}Cl/^{10}Be ratio (which decreases with an "effective" radioactive decay half-life of 375,000 years) would provide a way of dating ice older than ~100,000 years. Some early results, however, demonstrated that the deposition ratio of these two isotopes was not constant [*Elmore et al.*, 1987; *Suter et al.*, 1987], presumably because of their different chemical behaviors in the atmosphere (there are several gaseous species associated with chlorine, whereas beryllium is rapidly attached to aerosol particles). The observation that the ^{36}Cl/^{10}Be ratios in the Arctic and Antarctic are different (our unpublished data), supports this idea. It was hoped, however, that if one averaged over a sufficiently long period at one location the ratio might become more constant. It was therefore thought that the ^{36}Cl/^{10}Be ratio might contribute to a better understanding of the age of the ice near the bottom of the GRIP core. Unfortunately, however, this potential is seriously compromised by the uncertainty associated with the dust-related ^{10}Be. Because of its chemistry, one does not expect ^{36}Cl to have a dust-associated component. As can be seen in Figure 8, the ratio for samples prepared with 0.45 μm filters (and thus excluding the dust component) remains within the range observed in the upper part of the core. In contrast, the ratio for samples processed with 45 μm filters (thus including the dust component) decreases strongly near the bottom of the core, suggestive of very old (>300 kyr) ice. A more definitive interpretation will require a better understanding of the role of ^{10}Be associated with dust [*Baumgartner et al.*, 1997].

Conclusions

Although the experimental data presented here are not yet complete and pose a number of problems, the results permit several useful observations.

1. It has been shown for the first time that a significant fraction of ^{10}Be in melted Arctic polar ice is retained by 0.45 μm filters. While the origin of this "dust"-associated component is not yet definitively established, its presence may seriously complicate interpretations of ^{10}Be profiles in such ice.

2. Based on analyses of only ~12% of the available samples in Holocene ice, we observe tentative evidence for both solar and geomagnetically modulated variations of ^{10}Be production. More definitive statements will require a continuous ^{10}Be profile for this period.

3. As observed in polar ice elsewhere, the first-order origin of ^{10}Be concentrations variations at Summit is the "dilution" effect, due to variable snow accumulation rates. This is true not only between major glacial-interglacial periods, but also over shorter interstadial periods (Dansgaard-Oeschger events). In contrast, the ^{10}Be data do not support the interpretation of rapidly varying accumulation (i.e., climate) during the last interglacial. They can, however, be used to help place limits on the origin of the ice in these events.

4. As observed in Antarctic cores, there is a period of enhanced ^{10}Be flux at ~40 kyr B.P., lasting ~1500 years, but showing significant structure. This ^{10}Be peak provides a precise time marker for correlating Arctic and Antarctic ice cores that may in the future be extended to continental and polar climatic records.

5. The distribution of ^{10}Be between ice and dust components in the bottom 100 m of the core appears unique and may thus be a useful parameter for identifying possible incursions of such ice at any other depth in the core.

6. The ^{36}Cl/^{10}Be ratio shows significant fluctuations (factor of 2) over the whole length of the core, supporting the idea of a different chemical behavior of these two isotopes in the atmosphere. Over the bottom ~500 m of the core, this ratio decreases steadily by a factor of 2 if one considers total ^{10}Be (ice + filter retained), suggesting very old (>300 kyr) ice, but remains constant if one considers only ^{10}Be from the ice.

7. A more complete and reliable interpretation of ^{10}Be from the GRIP core will require a continuous profile and a better understanding of the dust problem.

Note added in proof. With a volcanic ash layer used as a marker, Vostok cores 3G and 4G are now believed to be offset ~4 m near the ^{10}Be peak (J.-R. Petit, private communication, 1997). This would place the ^{10}Be peak position on the Vostok deuterium curve in Figure 7 ~400 years earlier, thus bringing the two climate records closer in phase.

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