Advances in Geosciences
Vol. 14: Solar Terrestrial (2007)
Eds. Marc Duldig *et al.*

^{10}Be CONCENTRATIONS IN SNOW AT LAW DOME, ANTARCTICA FOLLOWING THE 29 OCTOBER 2003 AND 20 JANUARY 2005 SOLAR COSMIC RAY EVENTS

J. B. PEDRO[*,†], A. M. SMITH[‡], M. L. DULDIG[§], A. R. KLEKOCIUK[§], K. J. SIMON[‡], M. A. J. CURRAN[†,§], T. D. VAN OMMEN[†,§], D. A. FINK[‡], V. I. MORGAN[†,§] and B. K. GALTON-FENZI[*,†]

[*]*Institute of Antarctic and Southern Ocean Studies*
University of Tasmania, Hobart, Tasmania 7005, Australia

[†]*Antarctic Climate and Ecosystems CRC*
Hobart, Tasmania 7005, Australia

[‡]*Australian Nuclear Science and Technology Organisation*
Menai, New South Wales 2234, Australia

[§]*Australian Antarctic Division*
Kingston, Tasmania 7050, Australia

Recent model calculations have attempted to quantify the contribution of major energetic solar cosmic ray (SCR) events to ^{10}Be production.[1,2] In this study we compare modeled ^{10}Be production by SCR events to measured ^{10}Be concentrations in a Law Dome snow pit record. The snow pit record spans 2.7 years, providing a quasi-monthly ^{10}Be sampling resolution which overlaps with the SCR events of 29 Oct 2003 and 20 Jan 2005. These events were calculated to increase monthly ^{10}Be production in the polar atmosphere (>65° S geomagnetic latitude) by ∼60% and ∼120% above the GCR background, respectively.[2] A strong peak in ^{10}Be concentrations (>4σ above the 2.7 y mean value) was observed ∼1 month after the 20 Jan 2005 event. By contrast, no signal in ^{10}Be concentrations was observed following the weaker 29 Oct 2003 series of events. The concentration of ^{10}Be in ice core records involves interplay between production, transport, and deposition processes. We used a particle dispersion model to assess vertical and meridional transport of aerosols from the lower stratosphere where SCR production of ^{10}Be is expected to occur, to the troposphere from where deposition to the ice sheet occurs. Model results suggested that a coherent SCR production signal could be transported to the troposphere within weeks to months following both SCR events. We argue that only the 20 Jan 2005 SCR event was observed in measured concentrations due to favorable atmospheric transport, relatively high production yield compared to the 29 Oct 2003 event, and a relatively high level of precipitation in the Law Dome region in the month following the event. This result encourages further examination of SCR signals in ^{10}Be ice core data.

1. Introduction

^{10}Be ($t_{1/2} = 1.5 \times 10^6$ y) is a cosmogenic radionuclide produced by the interaction of cosmic rays with the Earth's atmosphere. The majority of ^{10}Be is produced by spalation of O, N, and Ar atoms by galactic cosmic rays (GCRs) within the energy range 0.1–10 GeV.[3] Following production, ^{10}Be is rapidly scavenged by aerosols and transported with air masses. Some ^{10}Be is deposited and archived within the annual snow layers of the polar ice caps. Ice core records of ^{10}Be concentration sample this archive and may be used to reconstruct the history of the factors which control the flux of cosmic rays to Earth, principally variations in solar activity[4,5] and variations in the geomagnetic field strength.[6,7]

Reconstructions of solar activity from ice core ^{10}Be concentrations can provide information for assessing solar forcing of climate in the past and hence the role of solar forcing in the present climate change.[8,9] However, reconstruction of reliable solar activity records from ^{10}Be has been limited by poor understanding of the influence of climate processes on ^{10}Be transport and deposition to the polar ice core sites.[5,10,11] The challenge in interpreting ^{10}Be records is to separate variations in production rate in the atmosphere caused by changes in solar activity from variations caused by atmospheric mixing, transport pathways, depositional processes, and changes in the precipitation rate.

A useful test of ^{10}Be records as proxies of solar activity is comparison of the ice core record to the instrumental record of solar activity and cosmic ray flux (CRF). Satellite records of solar activity were initiated in the 1970s. Instrumental records of CRF began with the installation of ionization chambers in the 1930s and then longer term, more stable ground-based neutron monitors in the 1950s. However, few comparisons between measured ^{10}Be and instrumental data exist.[12–14]

High-resolution ^{10}Be records spanning the modern era are also required to answer questions regarding possible solar cosmic ray (SCR) contribution to ^{10}Be production. SCRs are energetic particles emitted by the Sun generally in the energy range 0.1–50 GeV. During a major SCR event, the flux of particles arriving at the Earth's atmosphere can increase by orders of magnitude.[1] Over decadal and longer timescales, the production of ^{10}Be by SCRs is probably small (1%–2%) compared to production by GCRs.[1] However, over annual and shorter timescales SCR events may lead to significant production of ^{10}Be.

Recent theoretical studies have been presented for SCR contributions to ^{10}Be production during the ∼70 y record of instrumental SCR data.[1,2]

Usoskin *et al.*[1] argue that major events can cause up to a doubling in annual polar production of ^{10}Be. Webber *et al.*[2] suggest a smaller magnitude of SCR contribution, finding at most ∼50% enhancement in polar production (for the large February 1956 event). The calculations of Webber *et al.*[2] may be more reliable as they use updated atmospheric yields based on latest cross sections and a broader spectral range of incident protons. Measurements of ^{10}Be concentration in polar ice cores for periods spanning major SCR events are required to test and compare with modeled production rates. Such a comparison requires consideration of the latitudinal gradient in ^{10}Be production rate and the poorly constrained influences of atmospheric transport and deposition processes on ^{10}Be fallout.

Since SCRs are generally of lower rigidity (momentum per unit charge) than GCRs, they largely have access restricted to the polar regions. Polar production rates are therefore much more sensitive to SCRs than global average production rates. For example, the February 1956 SCR event was calculated to cause a ∼50% increase in polar production but only ∼12% increase in global average production.[2] It follows that the sensitivity of polar ice core records to SCR signals is highly dependent on the extent of meridional mixing of ^{10}Be produced over the polar regions. There is evidence that meridional mixing of ^{10}Be produced over the poles is limited, for example, from general circulation model calculations[15] and from comparison of ice core ^{10}Be concentrations with the amplitude of the geomagnetic and solar variations in the past.[12,16]

The atmospheric lifetime of ^{10}Be produced in the stratosphere is several years, whereas in the troposphere ^{10}Be is washed out by precipitation processes within several weeks.[17] Stratospheric concentrations of ^{10}Be are therefore enhanced with respect to the troposphere, and the temporally and spatially variable processes of stratosphere–troposphere exchange are of high relevance to ^{10}Be fallout. In addition to mixing processes, local effects including precipitation rate and frequency also influence the ^{10}Be fallout.[5,11,15]

In November 2005, a 4 m deep snow pit was extracted at Law Dome Summit (the "DSS0506 snow pit"). The snow pit has time resolution and dating accuracy at the quasi-monthly level and spans 2.7 y (March 2003 to November 2005). The record overlaps with two major SCR events: 29 Oct 2003 and 20 Jan 2005. Here, we examine the DSS0506 snow pit for SCR signals from these events. We argue that high temporal resolution ice core records of ^{10}Be concentration for periods spanning major SCR events

have potential to inform on both the magnitude of the SCR contribution to ^{10}Be production and the extent of latitudinal mixing of ^{10}Be. Through consideration of SCR production signals as "pulse" inputs of ^{10}Be to the atmosphere, information may also be obtained on atmospheric residence time and temporal variations in ^{10}Be transport.

2. Methods

2.1. *Sample site and characteristics*

Law Dome, East Antarctica (66° 46′ S, 112°48′ E) (Fig. 1) is well suited to the extraction of precisely dated and continuous chemical records in fine detail.[18,19] It experiences high snow accumulation (0.68 m y^{-1} ice equivalent) which provides for high temporal resolution records (quasi-monthly), low annual average wind speed (8.3 ms^{-1}), which minimizes surface disturbance and perennially low temperature (summer mean, −12.6°C) which precludes alteration of records by summer melt.[20]

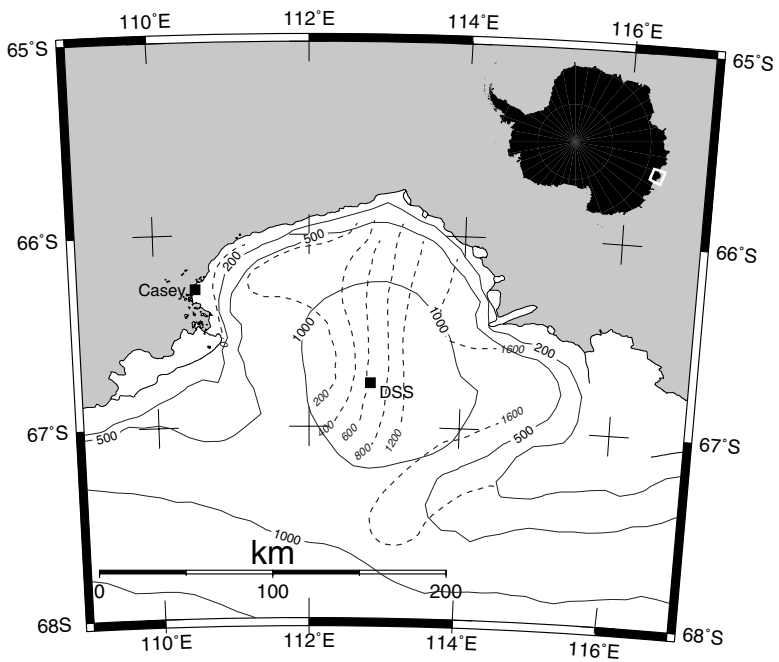

Fig. 1. Location of Dome Summit South (DSS) sample site, Law Dome, Antarctica with accumulation isopleths (mm ice equivalent, dashed lines) labeled.

2.2. *Sample collection*

The DSS0506 snow pit was excavated to a depth of 4.0 m in November 2005 near Law Dome Summit (1379 m asl, 66°46′ 11″ S, 112°48′ 41″ E). Forty 10 cm thick contiguous samples of approximately 1.5 kg mass were taken down the face of the snow pit by insertion of a 10 cm high × 20 cm×20 cm stainless steel scoop (following Pedro *et al.*[11]). Samples were immediately sealed in zip lock bags. To minimize contamination, all equipments were pre-rinsed with >18 MΩ cm^{-1} Milli-Q water, and personnel were equipped with dust masks and gloves.

2.3. *Sample preparation and measurement*

Sample preparation and accelerator mass spectrometer (AMS) measurements were carried out at the Australian National Tandem Accelerator for Applied Research (ANTARES) AMS using the techniques described in Child *et al.*[21] and Fink *et al.*[22] ^{10}Be concentrations were normalized to the National Institute of Standards (NIST) SRM 4325 ^{10}Be standard with an adjusted ratio of 3.02×10^{-11}.[23] Measurements of the NIST standard were reproducible within ±2%. Chemistry procedural blanks had very low values of ^{10}Be/^{9}Be ($<10 \times 10^{-15}$), indistinguishable from the carrier, indicating neither machine background nor chemistry processes introduced ^{10}Be at any significant level. After normalization to the NIST standard, error-weighted mean ^{10}Be/^{9}Be ratios ranged from (210–966)×10^{-15}, with overall errors of <4%.

The low ^{10}Be/^{9}Be ratios achieved in chemistry blanks from this study resulted from our use of a beryllium carrier derived from the mineral beryl which is very low in ^{10}Be, along with a step in our chemistry processing technique introduced specifically to remove boron (^{10}B) contamination from samples (^{10}B is a problematic isobaric interferent in AMS measurement of ^{10}Be). This technique involved fuming BeO from each sample with hydrofluoric acid in sulfuric acid to volatilize boron as BF_3 (K. Simon, in prep).

2.4. *Sample dating*

Fractionation of water isotopes during evaporation, transport, and precipitation processes leads to a well-established relationship between the oxygen isotope ratio (δ^{18}O) of precipitation and site temperature at high latitude sites.[24] This is the basis for climatic temperature reconstructions

from ice cores and is clearly evident in strong seasonal cycles in $\delta^{18}O$ observed at Law Dome.[25] The clear summer maxima in $\delta^{18}O$ provide dating horizons for Law Dome ice core records.

$\delta^{18}O$ samples were taken from the bulk melt water of each 10 cm resolution ^{10}Be snow pit sample. The snow pit $\delta^{18}O$ record is therefore registered directly with the ^{10}Be record. The timescale was determined by picking summer peaks in $\delta^{18}O$, assisted by interlocking with $\delta^{18}O$ records from other recent ice cores from Law Dome Summit and by reference to absolute dating tie points. Figure 2 shows the DSS0506 snow pit $\delta^{18}O$ record with $\delta^{18}O$ records from two recent ice cores, the DSS0506 ice core (drilled 300 m away from the DSS0506 snow pit) and the DSS0405 ice core (drilled at the same site two years earlier). The three records are generally in good agreement, which allows confidence in dating the DSS0506 snow pit. Some differences occur in the records due to local influences on snow removal, ablation, and accumulation. The separation of the DSS0506 ice core record and the DSS0506 snow pit record over the first 30 cm (ice equivalent) is due to the compression of the topmost portion of the snow pack by the drilling process and does not reduce confidence in dating.

Experimental work shows that on average $\delta^{18}O$ peaks on 10 January at Law Dome,[25] although year-to-year variations in timing of precipitation and annual temperature maximum are acknowledged. Year boundaries are labeled in Fig. 2 by attribution of 10 January to the annual $\delta^{18}O$ maxima. Where there is disagreement between the three $\delta^{18}O$ records, a mean is taken. Firm tie points on the dating scale are the sample dates for the DSS0506 snow pit (8 Nov 2005) and the drilling date of the DSS0405 ice core (31 Oct 2004).

The dates of samples between summer peaks are interpolated assuming even snow accumulation over the course of a year. Evidence suggests that this assumption of uniformly distributed precipitation is valid when averaged over many years,[25] however in any given year, precipitation biases are seen.[19] The bottom date for the record is determined by interlocking the snow pit $\delta^{18}O$ record with the DSS0506 ice core $\delta^{18}O$ record. Our best estimate of error associated with dating of the snow pit record is ± 1 month, with dating confidence firmer in the vicinity of the 8 Nov 2005 and 31 Oct 2004 tie points.

Reproducibility of snow pit records extracted from Law Dome is expected to be good; prior studies found that trace chemical records extracted from multiple snow pits on Law Dome provide records which are in agreement over at least a 12 km transect.[26]

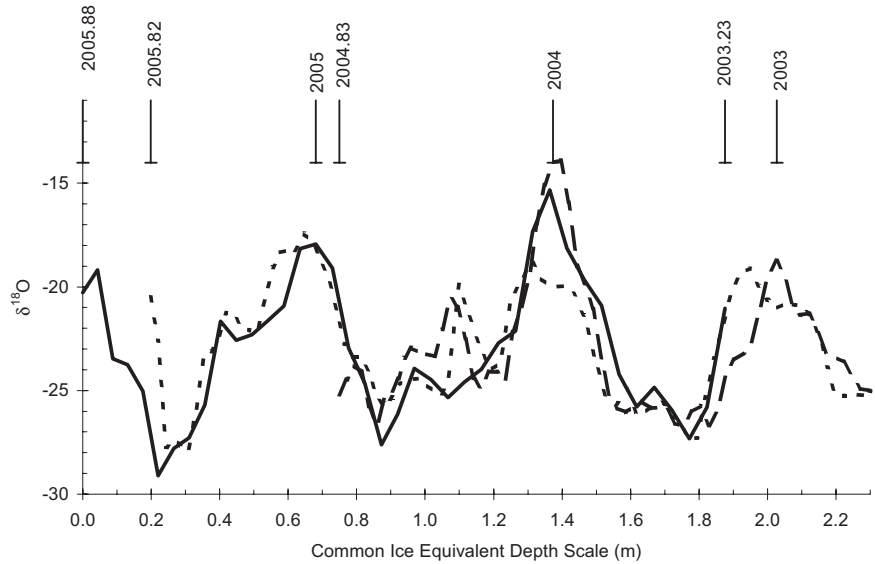

Fig. 2. Dating of the DSS0506 snow pit record was carried out using δ^{18}O with cross reference to two recent ice core records from Law Dome and registration against firm dating tie points. DSS0506 snow pit δ^{18}O record (solid line), DSS0506 ice core δ^{18}O record (dashed line), and DSS0405 ice core δ^{18}O record (dotted line). The interpolated year boundaries for 2005, 2004, and 2003 are labeled. Firm dating tie points are sampling dates of the DSS0506 snow pit (2005.88), DSS0506 ice core (2005.82), and DSS0405 ice core (2004.83). The bottom date of the DSS0506 snow pit is interpolated (2003.23). The date scale is expected to be accurate to ±1 month with more confidence close to dating tie points.

3. Results

3.1. ^{10}*Be concentration and variability in the DSS0506 snow pit*

The mean (and SD) of the concentration of ^{10}Be in the 40 snow pit samples was $(6.0 \pm 2.0) \times 10^3$ atoms g^{-1}. This value is similar to earlier measurements of ^{10}Be concentration in modern Law Dome ice, e.g. 5.6×10^3 atoms g^{-1} reported by Pedro *et al.*[11] and 7.6×10^3 atoms g^{-1} reported by Smith *et al.*[27] Figure 3(a) shows variation in ^{10}Be concentration with time for the snow pit record. The record spans March 2003 to December 2005 with dating accuracy estimated at ±1 month. The range of concentrations varies by up to fourfold throughout the record from 2.7 to 12.6×10^3 atoms g^{-1} (with 4% analytical errors). Concentrations are generally higher during the late

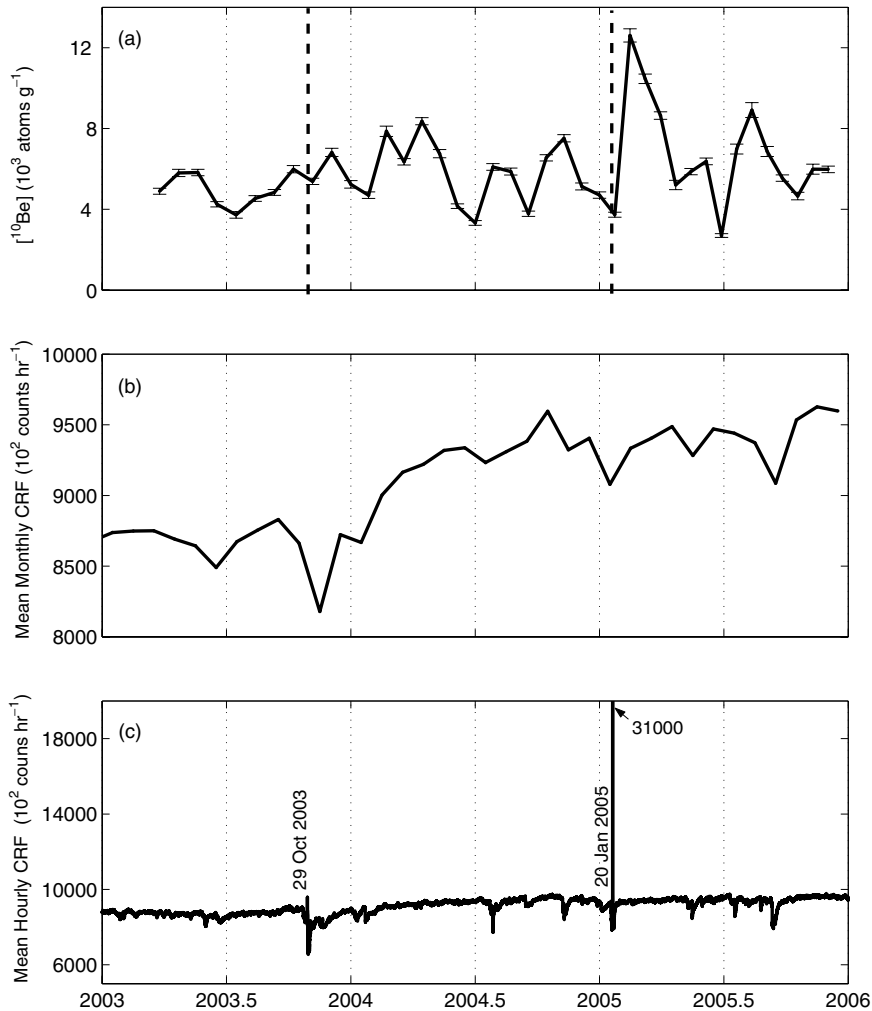

Fig. 3. (a) ^{10}Be concentration in the DSS0506 snow pit showing timing of 29 Oct 2003 and 20 Jan 2005 SCR events (dashed lines); (b) McMurdo monthly mean CRF[31]; (c) McMurdo hourly mean CRF[31] with the 29 Oct 2003 and 20 Jan 2005 Ground Level Enhancements labeled.

austral summer (summer) and lower during the austral winter (winter). This is in agreement with a prior snow pit study at Law Dome which found a late summer maximum in ^{10}Be concentration.[11] Recurrence of the late summer maximum in ^{10}Be concentrations suggests that atmospheric

transport is conducive to increased ^{10}Be fallout at this time of year; prior studies have suggested that an enhancement in stratosphere to troposphere exchange may be implicated.[11,28,29]

Over time periods of less than 1 year, meteorological processes are expected to dominate variability in ^{10}Be concentrations.[10] However, some secondary influence by production rate changes in the atmosphere may also be imprinted on the record. The most prominent feature in the snow pit record is the strong increase in ^{10}Be concentration during February of 2005. A possible SCR origin for this peak is discussed in Sec. 3.3.

3.2. ^{10}*Be concentration and monthly mean cosmic ray flux*

Solar modulation of the GCR flux arriving at the Earth's atmosphere over the course of a 11-yr Schwabe cycle is calculated to cause variation in global ^{10}Be production between 15% and 50%.[30,31] At polar latitudes ^{10}Be production is more sensitive to variation in the GCR flux; it has been calculated that production changes over recent Schwabe cycles at polar latitudes are on average close to 80%.[2] Here we examine the possible influence of shorter term (monthly) variations in the CRF as recorded by ground-based neutron monitors on ^{10}Be concentrations in ice.

Figure 3(b) shows monthly mean CRF data[32] recorded at the McMurdo neutron monitor (the closest monitor to Law Dome with available data). This may be compared to ^{10}Be concentration in the snow pit in Fig. 3(a). No correlation is apparent between the monthly CRF data and quasi-monthly ^{10}Be concentrations, either direct or delayed. Over the period of the snow pit record, mean monthly CRF varies by at most ∼5%, and such relatively small variations in CRF cannot explain the large variations in ^{10}Be concentration observed in the snow pit. Our data support the contention that meteorological processes are the principal cause of variation in ice core concentrations of ^{10}Be over time periods of <1 y (e.g. Beer[10]). To our knowledge this is the first comparison of monthly resolution CRF data to ^{10}Be concentrations at similar resolution from a polar ice core site.

We note that the positive trend in CRF over the period spanned by the record is matched by an increasing trend in ^{10}Be concentration. This is consistent with the Schwabe cycle influence on CRF and ^{10}Be production in the atmosphere, but the level of variability in the ^{10}Be concentration precludes affirmation of a casual link.

3.3. ^{10}*Be concentration and solar cosmic ray events*

^{10}Be production is highly sensitive to the energy spectrum and particle distribution of individual SCR events. Webber *et al.*[2] have compiled new data sets of both integral and differential fluence spectra for major SCR events since 1940 and used these to calculate ^{10}Be production for each event.

Two significant SCR events occurred within the period covered by the DSS0506 snow pit record; these are 29 Oct 2003 and 20 Jan 2005. Integrated particle fluences and calculated polar production of ^{10}Be for these events are listed in Table 1. 29 Oct 2003 consisted of a series of events that were among the strongest in the current solar cycle (29 Oct 2003 events). The 20 Jan 2005 event exhibited high particle intensity in the higher energy region of the spectrum (i.e. a flatter spectrum) (20 Jan 2005 events). Both events caused ground-level enhancements (GLEs) in CRF recorded by neutron monitors, though the 29 Oct 2003 events had much smaller amplitudes. At peak intensity, the 20 Jan 2005 event resulted in a 2000% increase in CRF to the polar regions,[2] making it the strongest GLE in over 50 years. ^{10}Be production for the 29 Oct 2003 events peaked at ∼100 MeV, whereas the flatter spectrum of the 20 Jan 2005 event moved the production peak to higher energies.[2]

In terms of ^{10}Be production, SCRs during the 29 Oct 2003 event were calculated to contribute an additional ∼5% (with respect to GCR-produced

Table 1. Calculated polar ^{10}Be production and particle fluences (F) for 29 Oct 2003 and 20 Jan 2005 SCR events.

Event	GLE peak intensity %[a]	Relative increase in annual ^{10}Be production[b]	Relative increase in monthly ^{10}Be production[c]	$F > 350$ MeV	$F > 100$ MeV	$F > 10$ MeV
29 Oct 2003	10, 35, 16	5%	60%	2.5×10^6	9.0×10^7	1.0×10^{10}
20 Jan 2005	2000	10%	120%	1.5×10^7	6.0×10^7	8.0×10^8

[a]Increases in peak intensity ("GLE" or Ground Level Enhancement) recorded by neutron monitors at sea level for the polar regions.
[b]The % SCR contribution to ^{10}Be production with respect to GCR-produced ^{10}Be. All % contributions are expressed relative to GCR production during 2003 (the year of the "reference event"; 29 Oct 2003, considered by Webber *et al.*[2]
[c]Relative monthly contributions are calculated by attributing all SCR production to the month of the SCR event. All % monthly contributions are expressed relative to mean monthly GCR production for 2003, assuming GCR production was evenly distributed over the year. Table adapted from Webber *et al.*[2]

^{10}Be) to polar production in 2003, or an additional ∼60% to polar production during October.[2] The 20 Jan 2005 event was calculated to contribute an additional ∼10% to polar production for 2005 or an additional ∼120% of polar production for January.[2]

Here we look for a concentration signal in the snow pit record related to these two SCR events. This work is greatly facilitated by the theoretical predictions of SCR contribution to polar production of ^{10}Be as estimated by Webber *et al.*[2]

Figure 3(c) shows hourly mean CRF data[32] recorded at the McMurdo neutron Monitor. The GLEs associated with 29 Oct 2003 and 20 Jan 2005 events show up clearly as spikes in CRF intensity (note the greater amplitude of the 20 Jan 2005 event). The relationship between measured ^{10}Be concentration in the DSS0506 snow pit (Fig. 3(a)) and the timing and amplitude of modeled ^{10}Be production (Table 1) is considered for each event in turn.

3.3.1. *29 Oct 2003 SCR events*

^{10}Be concentrations do not show any significant deviation from the mean following the 29 Oct 2003 events (Fig. 3(a)). This may be partly due to the smaller amplitude of the production increase associated with the event (approximately half the amplitude of 20 Jan 2005 according to Webber *et al.*[2]). However, we expect the principal reason is that meteorological influences in the 1–2 months following the event are not conducive to rapid transport of a production signal to the ice core record.

3.3.2. *20 Jan 2005 SCR event*

A sharp peak in ^{10}Be concentration is observed following the 20 Jan 2005 event (Fig. 3(a)). ^{10}Be concentrations in the two samples following the event were 12.6 and 10.5×10^3 atoms g^{-1} (with 4% analytical errors); the mean for these samples is $>4\sigma$ above the mean for the remainder of the data set. The ^{10}Be peak was dated to mid-February 2005, indicating a delay of ∼1 month between the SCR event and the maximum in ^{10}Be concentration. Our dating of the concentration peak is reliable and supported by meteorological data from nearby Casey station and prior work.[25,33] Significantly, there was precipitation throughout the summer period, confirming that the δ^{18}O record used in dating (refer to Sec. 2.4) was reliable for the period that included the event. Furthermore, the monthly mean temperature record for Casey confirms that the annual temperature maxima occurred in January

2005, consistent with the timing of the maximum in $\delta^{18}O$ assumed in dating (refer also to Sec. 4.4).

The length of delay between the production signal and concentration signal in the ice core record depends on atmospheric mixing and deposition processes. We note that a delay of ∼1 month is similar to the reported tropospheric residence time of ^{10}Be.[17] The approximate doubling in ^{10}Be concentration following the event is consistent with an approximate doubling in ^{10}Be production caused by the SCR event for January. Hence, the timing and amplitude of the ^{10}Be peak are consistent with a connection to the 20 Jan 2005 event. However, variable influences of meteorological effects on the record make it difficult to ascertain causality. We note that the peak occurs in summer when there is evidence that ^{10}Be arrival to polar sites is augmented by meteorological effects. Meteorological conditions at the time of the two SPE events are now discussed.

4. Discussion

4.1. *Atmospheric transport and deposition of SCR-produced ^{10}Be*

In the polar regions ^{10}Be production dominantly occurs in the stratosphere; for example, Masarik and Beer[30] calculate that stratospheric production of ^{10}Be over the poles may account for up to eight times tropospheric production. In addition, the energies and fluxes of energetic particles entering the atmosphere following large solar events are such that significant short-timescale (typically 10 s of minutes to hours) enhancement of spalation products takes place preferentially in the lower stratosphere at polar latitudes.[2]

Following production, ^{10}Be is rapidly scavenged by aerosol particles.[17] In the stratosphere, the aerosol population is primarily comprised of sulphate particles, with a small admixture of meteoritic material.[34,35]

The concentration of ^{10}Be in ice core records involves interplay between production, transport, and deposition processes. We propose that observation of enhanced ^{10}Be concentrations in ice at Law Dome subsequent to SCR events (i.e. a SCR production signal) requires (a) that vertical transport permits polar stratospheric aerosols tagged with ^{10}Be to mix down into the troposphere before meridional mixing has damped out the production signal in the atmosphere, and (b) that precipitation processes are favorable for scavenging aerosols tagged with ^{10}Be and depositing them

to the ice sheet. The amplitude of the SCR production signal that is sufficient to be observed in the ice core record will depend on the efficacy of these processes. We assess the two SCR events in the snow pit record with respect to these criteria using dispersion modeling of stratospheric aerosol pathways, empirical data on stratospheric aerosol arrival to the Antarctic ice sheet, and meteorological data on snow deposition in the Law Dome region (Secs. 4.2–4.4).

4.2. *Dispersion modeling of stratospheric aerosol pathways*

To assess the extent of meridional and vertical transport of the production signal for each SCR event, we used the dispersion component of the Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT) model,[36] and performed forward analysis of simulated aerosol pathways. A population of approximately 2400 spherical particles of density $1.7\,\mathrm{g\,cm^{-3}}$ and $0.5\,\mu$m diameter (to simulate sulphate aerosols within the size range reported to predominantly scavenge ^{10}Be (McHargue and Damon[17]) were released from 41 sites spread evenly on a 1000 km grid south of $\sim$60° S. Particles were released over a period of 5 h approximating the duration of the peak solar particle fluxes for the 20 Jan 2005 and 29 Oct 2003 SCR events. Different release heights were trialed consistent with the expected height of ^{10}Be production.[2] The HYSPLIT model was used to trace the location of particles at 24 h intervals until subsidence to the troposphere occurred (subsidence to troposphere was judged with reference to NCEP/NCAR reanalysis data and using the World Meteorological Organisation thermal tropopause definition). Simulations spanned 90 days of simulated time.

Results from the dispersion modeling suggested significant aerosol subsidence to the troposphere within weeks to months after particle release for all simulations. This is rapid with respect to the 1- to 2-year mean stratospheric residence time of ^{10}Be given by Koch and Mann[37] and Dibb *et al.*[38] Results suggest slightly stronger subsidence for the 29 Oct 2003 events. However, a coherent signal was more likely to be transmitted for ^{10}Be produced in the lowermost part of the stratosphere (200 hPa level, $\sim$12.5 km altitude), and the higher intensity of high-energy particles associated with the 20 Jan 2005 event (Table 1) is expected to cause greater production in this region compared to the 29 Oct 2003 event.

The modeled aerosol descent would be expected to transmit a SCR production signal to the troposphere for both SCR events. Support for the relatively rapid aerosol descent interpreted from our model results is

offered by a recent global circulation model experiment, which suggested incomplete meridional mixing of ^{10}Be produced over the poles.[15] A detailed consideration of modeled stratospheric aerosol dispersion following the SCR events will be presented in a separate manuscript (Klekociuk *et al.,* in prep).

4.3. *Empirical data on arrival of stratospheric aerosols to polar regions*

An important difference between the meteorological conditions over Antarctica for the two events is the condition of the polar vortex. The polar vortex usually forms in late May and persists until between October and December. Atmospheric dynamics within the vortex are dependent on the altitude and latitude under consideration, and exhibit considerable interannual variability.[39] In general, while the vortex is present, temperature gradients across the tropopause are weak, and enhanced subsidence of stratospheric air is expected.[40] Consideration of vortex area data (for example, vortex area analyses for the 450 K potential temperature surface in the lower stratosphere[41]) shows that the 29 Oct 2003 SCR occurred while the vortex was breaking down and the 20 Jan 2005 event occurred well after the vortex of the 2004 winter had dissipated. One might expect that enhanced subsidence of stratospheric air while the vortex is present would promote the arrival of stratospheric aerosols to the polar regions. However, numerous empirical data suggest that stratospheric aerosols predominantly arrive to the Antarctic ice sheet in late summer. For example, ground-based measurements of the stratospheric aerosol markers ^{7}Be/^{10}Be ratio and ^{7}Be/^{210}Pb ratio[42] show late summer maximum at Neuymaer[29,43] as do ^{7}Be levels at South Pole and coastal Terre Adélie.[42,44,45] ^{10}Be concentrations in snow also show evidence of enhancement during late summer and autumn, for example at Law Dome[11] and Dye 3, South Greenland.[28] No mechanism has been clearly demonstrated to explain the summer to autumn maximum in stratospheric aerosol arrival to polar regions. Wagenbach[29] suggested that increased vertical mixing may occur at this time due to weakening of the tropospheric surface temperature inversion, whereas Beer *et al.*[28] proposed that a summer to autumn peak may be linked to longer-range transport of stratospheric aerosol from mid-latitude tropospheric injections.

Our aerosol dispersion results suggest that meridional mixing following the two SCR events was probably insufficient to damp out the production signals. Empirical data from aerosol and ice core measurements of ^{10}Be

suggests conditions would be more favorable for the arrival of stratospheric aerosols to the Law Dome in summer following the 20 Jan 2005 SCR rather than the 29 Oct 2003 event. Deposition processes are now considered.

4.4. *Deposition processes*

Incorporation of ^{10}Be into polar ice sheets is by a combination of dry fallout of ^{10}Be containing aerosols ("dry" deposition) and scavenging of ^{10}Be containing aerosols by precipitation ("wet" deposition).[46] The relative contribution of wet to dry precipitation is site- and climate-specific.[15] Studies of ^{10}Be deposition at Law Dome have found that wet deposition, i.e. the arrival of aerosol particles with precipitation is dominant.[11,27] For this site, registration of a concentration signal in high-resolution ^{10}Be data will be more likely if significant aerosol scavenging by precipitation occurred in the period following the SCR event.

We consulted precipitation records from the Australian Casey Station, the nearest meteorological station to Law Dome (∼100 km away). Prior inter-comparison of snow pit, ice core, and meteorological data from Law Dome and Casey station has demonstrated that meteorological observations at Casey can inform on meteorological conditions in the Law Dome region.[25,33] Precipitation in February 2005 following the 20 Jan 2005 event was high (22.8 mm), 65% above the 30-year mean, and 70% of the precipitation at Casey during the 3-month period spanning the event (December, January, February) arrived in February. By contrast, the precipitation in the month following the 29 Oct 2003 events was low (12 mm), slightly below the 30-year mean, and only 30% of the precipitation at Casey during the 3-month period spanning the events arrived in November. These data suggest that aerosol scavenging by precipitation is likely to have been facilitated in the period following the 20 Jan 2005 event and not so well facilitated following the weaker 29 Oct 2003 events.

5. Summary and Conclusions

Measured ^{10}Be concentrations in the 2.7-y DSS0506 snow pit were compared with CRF data from the McMurdo neutron monitor and with modeled ^{10}Be productions from SCR events. There was no relationship observed between measured ^{10}Be concentration and monthly mean CRF data. This result confirms that over short (sub-annual to annual) timescales, ^{10}Be concentration in ice core records is mainly influenced by meteorological

effects. However, we find evidence that the sharp production signals associated with SCR events may be imprinted on ^{10}Be concentration.

We investigated the possible connection of ^{10}Be concentration and modeled the ^{10}Be production for the SCR events of 29 Oct 2003 and 20 Jan 2005. A strong peak in concentration was observed ∼1 month after the 20 Jan 2005 event but no increase in concentration was observed following the 29 Oct 2003 event. We argued that atmospheric conditions and deposition processes following the 20 Jan 2005 event were likely to be favorable for registration of the production signal, whereas conditions following the 29 Oct 2003 events were less favorable. Aerosol dispersion modeling suggested that for both events ^{10}Be aerosols tagged with ^{10}Be reached the troposphere before excessive smoothing of the production signal occurred. The 20 Jan 2005 event exhibited high fluence in the high-energy part of the spectrum; hence, production was expected to have occurred lower in the stratosphere from where mixing into the troposphere was more efficient. Furthermore, empirical data from ice core and aerosol measurements at Antarctic stations show that stratospheric aerosols are preferentially delivered to the polar region in the late summer, which corresponds with the observed February maximum in ^{10}Be concentration from the 20 Jan 2005 event. Finally, precipitation data from nearby Casey station shows that precipitation was particularly high in February following the 20 Jan 2005 event and suitable for leaching of the production signal from the atmosphere, as opposed to November following the 29 Oct 2003 SCR when precipitation was low. Considering these transport and deposition factors along with the greater amplitude of ^{10}Be production by the 20 Jan 2005 SCR, it seems reasonable that only the 20 Jan 2005 SCR was observed in ^{10}Be concentrations. It should be noted, however, that we cannot rule out that the strong peak in concentration during February 2005 has some other cause.

Identification of production signals in ^{10}Be data is complex due to the sensitivity of ^{10}Be concentration to variable climate and meteorological processes. A detailed analysis over a longer record is required to constrain the link between measured ^{10}Be concentrations and SCR production of ^{10}Be. We plan to carry out such an analysis using the most recent 60 years of the high-resolution DSS0506 ice core record.

Acknowledgments

This work was assisted by the Australian Government's Cooperative Research Centres Programme, through the Antarctic Ecosystems and

Climate Cooperative Research Centre (ACE CRC). The Australian Antarctic Division supported the field campaign to sample the DSS0506 snow pit and ice core records. The Cosmogenic Climate Archives of the Southern Hemisphere (CcASH) project of the Australian Nuclear Science and Technology Organisation (ANSTO) supported ^{10}Be measurements. Joel Pedro is grateful for research funding from an Australian Institute of Nuclear Science and Engineering Postgraduate Research Award and an Australian Postgraduate Award. The Bartol Research Institute neutron monitor program is supported by National Science Foundation grant ATM-0527878.

References

1. I. G. Usoskin, S. K. Solanki, G. A. Kovalstsov, J. Beer and B. Kromer, *Geophys. Res. Lett.* **33** (2006) L08107.
2. W. R. Webber, P. R. Higbie and K. G. McCracken, *J. Geophys. Res.* **112** (2007) A10106, doi: 10.1029/2007JA012499.
3. D. Lal and B. Peters, *Handbuch der Physik* **46** (1967) 551.
4. G. M. Raisbeck and F. Yiou, in *The Ancient Sun: Fossil Record in the Earth, Moon and Meteorites; Proceedings of the Conference*, eds. R. O. Peppin, J. A. Eddy and R. B. Merill (Pergamon Press, New York and Oxford, 1980), pp. 185–190.
5. R. Muscheler, F. Joos, J. Beer, S. A. Muller, M. Vonmoos and I. Snowball, *Quart. Sci. Rev.* **26** (2007) 82–97.
6. G. Wagner, J. Masarik, J. Beer, S. Baumgartner, D. Imboden, P. W. Kubik, H.-A. Synal and M. Suter, *Nucl. Instrum. Methods, Phys. Res. Sect. B* **172**(1–4) (2000) 597.
7. R. Muscheler, J. Beer, P. W. Kubik and H. A. Synal, *Quart. Sci. Rev.* **24**(16–17) (2005) 1849.
8. E. Bard and M. Frank, *Earth Planet. Sci. Lett.* **248** (2006) 1.
9. C. M. Ammann, F. Joos, D. S. Schimel, B. L. Otto-Bliesner and R. A. Tomas, *Proc. Nat. Acad. Sci. USA* **104**(10) (2007) 3713.
10. J. Beer, *Space Sci. Rev.* **94** (2000) 53.
11. J. Pedro, T. van Ommen, M. Curran, V. Morgan, A. Smith and A. McMorrow, *J. Geophys. Res.* **111**(D10) (2006) D21105, doi: 10.1029/2005JD006764.
12. E. J. Steig, P. J. Polissar, M. Stuiver, R. C. Finkel and P. M. Grootes, *Geophys. Res. Lett.* **25** (1996) 523.
13. A. Aldahan, G. Possnert, S. J. Johnsen, H. B. Clausen, E. Isaksson, W. Karlen and M. Hansson, *Proc. Ind. Acad. Sci.* (*Earth Planet. Sci. Lett.*) **107**(2) (1998) 139.
14. H. Moraal, R. Muscheler, L. du Piessis, P. W. Kubik, J. Beer, K. G. McCracken and F. B. McDonald, *S. Afr. J. Sci.* **101** (2005) 299.
15. C. V. Field, G. A. Schmidt, D. Koch and C. Salyk, *J. Geophys. Res.* **111** (2006) D15107, doi: 10.1029/2005JD006410.

16. K. G. McCracken, *J. Geophys. Res.* **109** (2004) A04101, doi: 10.1029/2003JA010060.
17. L. R. McHargue and P. E. Damon, *Rev. Geophys.* **29**(2) (1991) 141.
18. M. A. J. Curran, T. D. van Ommen and V. I. Morgan, *Ann. Glaciol.* **27** (1998) 385.
19. A. J. McMorrow, T. D. van Ommen, V. Morgan and M. A. J. Curran, *Ann. Glaciol.* **39** (2004) 34.
20. V. I. Morgan, C. W. Wookey, J. Li, T. D. van Ommen, W. Skinner and M. F. Fitzpatrick, *J. Glaciol.* **43** (1997) 3.
21. D. Child, G. Elliott, C. Misfud, A. M. Smith and D. Fink, *Nucl. Instrum. Methods Sect. B* **172** (2000) 856.
22. D. Fink, B. McKelvey, D. Hannan and D. Newsome, *Nucl. Instrum. Methods Sect. B* **172** (2000) 838.
23. D. Fink and A. Smith, *Nucl. Instrum. Methods Sect. B* **259** (2007) 600.
24. W. Dansgaard, *Tellus* **16** (1964) 436.
25. T. D. van Ommen and V. Morgan, *J. Geophys. Res.* **102**(D8) (1997) 9351.
26. A. J. McMorrow, M. A. J. Curran, T. D. van Ommen, V. Morgan and I. Allison, *Ann. Glaciol.* **35** (2002) 463.
27. A. M. Smith, D. Fink, D. Child, V. A. Levchenko, V. Morgan, M. Curran and D. Etheridge, *Nucl. Instrum Methods Sect. B* **172** (2000) 847.
28. J. Beer *et al.*, *Atmos. Environ.* **25A**(5/6) (1991) 899.
29. D. Wagenbach, in *Chemical Exchange Between the Atmosphere and Polar Snow*, eds. E. W. Wolff and R. C. Bales, NATO ASI Series, Vol. I 43 (Springer-Verlag, Berlin Heidelberg, 1996), pp 173–199.
30. J. Masarik and J. Beer, *J. Geophys. Res.* **104**(D10) (1999) 12099.
31. W. R. Webber and P. R. Higbie, *J. Geophys. Res.* **108**(A9) (2003) 1355, doi: 10.1029/2003JA009863.
32. Bartol Research Institute, Neutron Monitor Data for McMurdo, http://neutronm.bartol.udel.edu/.
33. A. J. McMorrow, PhD thesis, University of Tasmania, Hobart, Australia, 2006.
34. R. C. Whitten, O. B. Toon and R. P. Turco, *Pageoph.* **118** (1980) 86.
35. D. M. Murphy, D. S. Thompson and M. J. Mahoney, *Science* **282** (1998) 1664.
36. R. R. Draxler, *J. Appl. Meteorol.* **42** (2003) 308.
37. D. M. Koch and M. E. Mann, *Tellus, Ser. B* **48** (1996) 387.
38. J. E. Dibb, D. L. Meeker, R. C. Finkel, J. R. Southon, M. W. Caffee and L. A. Barrie, *J. Geophys. Res.* **99**(12) (1994) 855.
39. G. L. Manney, R. W. Zurek, A. ONeill and R. Swinbank, *J. Atmos. Sci.* **51**(20) (1994) 2973.
40. W. Schwerdtfeger, in *Weather and Climate of the Antarctic* (Elsevier, New York, 1984), p. 261.
41. NOAA, National Weather Service, National Centre for Environmental Prediction, Climate Prediction Centre, Time Series of the Southern Hemisphere Polar Vortex at 450K, http://www.cpc.ncep.noaa.gov/products/stratosphere/polar/polar.shtml.

42. G. M. Raisbeck, F. Yiou, M. Fruneau, J. M. Loiseaux, M. Lieuvin and J. C. Ravel, *Geophys. Res. Lett.* **8** (1981) 1015.
43. D. Wagenbach, U. Goerlach, K. Moser and K. O. Muennich, *Tellus, Ser. B* **40** (1988) 426.
44. W. Maenhaut, W. H. Zoller and D. G. Coles, *J. Geophys. Res.* **84** (1979) 3131.
45. J. Sanak, G. Lambert and B. Ardouin, *Tellus, Ser. B* **37** (1985) 109.
46. G. M. Raisbeck and F. Yiou, *Ann. Glacio.* **7** (1985) 138.