Meteoric Be-10 from Sirius Group suggests high elevation McMurdo Dry Valleys permanently frozen since 6 Ma

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\section*{A B S T R A C T}

A long-standing debate concerning Neogene Antarctic climate in the McMurdo Dry Valleys relies largely on evidence from landscape evolution, glacial modeling and stratigraphy. We provide new evidence from meteoric \(^{10}\)Be for the onset of frozen, hyper-arid conditions on a high elevation (1840 m) interfluve at Table Mountain. A simple decay model for the co-occurrence of meteoric \(^{10}\)Be and illuviated clay in cores of ice-cemented glacial sediments indicates that the clays were actively migrating down from the surface in a warm climate until the system froze between 6 and 9 Ma. Although this age range may be sensitive to possible interference by in situ produced \(^{10}\)Be, the implied minimum age of 6 Ma for the Sirius Group indicates that the Dry Valleys were permanently frozen down to this elevation at this time. The model also suggests denudation rates of 1–6 cm Myr\(^{-1}\) since freezing. These data provide an independent test of glacial-stratigraphic evidence used to determine Antarctic paleoclimate.

\section*{1. Introduction}

Evidence for paleoclimate over the past 15 million years in the McMurdo Dry Valleys (MDV), Antarctica, seems contradictory. A large body of stratigraphic and glacial geomorphic evidence (Denton et al., 1993; Marchant et al., 1996; Sugden et al., 1995) indicates that at high (> 1000 m) elevations in the MDV a cold, dry polar climate has persisted since ca. 12 Ma (Lewis et al., 2007). ANDRILL cores show that the West Antarctic Ice Sheet was greatly reduced in the Ross Embayment and indicate numerous periods of open-water conditions during warmer-than-present interglacials between 11 and 1 Ma (McKay et al., 2009).

In this paper, we attempt to reconcile this coastal warmth with the high elevation polar conditions by applying a new method, which is independent of landscape evolution, glacial modeling and stratigraphy. We use meteoric \(^{10}\)Be, measured to depths of 4 m in cores from ice-cemented sediments of the Sirius Group and an adjacent debris flow at Table Mountain (Fig. 1). The presence of meteoric \(^{10}\)Be is significant because, in the present MDV environment, \(^{10}\)Be must enter the soil profile attached to particles from the surface. Illuviated clay, which has been recognized in thin sections of our samples, offers a mechanism for emplacement of this \(^{10}\)Be.

The translocation and illuviation of clay particles from the surface into deeper soil horizons is well documented in temperate and semiarid environments (Birkeland, 1999; Buurman et al., 1998). This process requires the downward percolation of water, which transports clay through an open-pore network until it falls out of traction because of blocked pathways or reduced flow. However, clay illuviation is problematic in the Transantarctic Mountains where rain has not fallen for millions of years (Marchant et al., 1996; Sugden et al., 1995). Modeling of \(^{10}\)Be decay in our profiles allows us to determine when \(^{10}\)Be was ‘closed off’ from the surface as the sediments became permanently frozen, and hence, the paleoenvironment under which the profiles developed.

\subsection*{1.1. Meteoric \(^{10}\)Be in the Dry Valleys}

Cosmic rays produce \(^{10}\)Be (half-life = 1.36 Myr; Nishiizumi et al., 2007) in the upper atmosphere, and particle-reactive \(^{10}\)BeO or \(^{10}\)Be(OH)\(_2\), attaches to aerosols, which are brought to the Earth’s surface mainly by wet precipitation (McHargue and Damon, 1991). This hydrolyzed \(^{10}\)Be rapidly adsorbs onto fine particles at the soil surface. Virtually all natural species of Be are insoluble in waters of pH > 4 (Takahashi et al., 1999), which are
typical for the MDV. Thus, the mobility of meteoric $^{10}$Be depends largely on the mobility of fine particles to which it is attached. At the Earth’s surface, the production rate of in situ $^{10}$Be, commonly used for exposure age dating, is roughly $10^3$ times less than the production of meteoric $^{10}$Be in the atmosphere (Gosse and Phillips, 2001; McHargue and Damon, 1991). Because of this, distinguishing these two types of Be is generally not a problem. However, at Table Mountain where altitude is high, erosion rates are slow, and soil surfaces may be as old as 12 Ma (Lewis et al., 2007), in situ $^{10}$Be cannot be ignored.

In this paper, we first present an age model that assumes meteoric $^{10}$Be in the soil without an in situ component. This is followed by a separate discussion on how a maximum in situ component might influence these results.

Meteoric $^{10}$Be has been found to depths of 1 m in several Dry Valley soils (Graham et al., 2002). In Wright Valley about 50 km north of Table Mountain, Schiller et al. (2009) showed that the transfer of $^{10}$Be into the ground has been inactive since deposition of the Hart Ash (320 m.a.s.l), 3.9 Ma. High concentrations of $^{10}$Be below the ash indicate clay illuviation in a warmer and wetter climate than today. When illuviation took place is not clear, but in this paper we use a two-profile method to derive a ‘closure age’ when the $^{10}$Be was sealed or frozen off from the atmosphere at a high elevation site.

2. Sample sites and methods

Cores were taken from the NW flank of Table Mountain (Fig. 1), where a thin linear band (2 x 5 km) of Sirius Group sediments overlies Beacon Supergroup sediments and intrusions of Ferrar Dolerite. The age of the Sirius Group has been hotly debated (Miller and Mabin, 1998), but stratigraphic relationships (e.g. Barrett and Powell, 1982; Denton et al., 1993), volcanic ash ages (Marchant et al., 1996) and exposure ages (Ivy-Ochs et al., 1995) strongly suggest that it is at least 15 Myr old in the MDV. Table Mountain is mantled by diamicts interpreted as debris flows, which are defined by areas of large-scale polygonal ground (Fig. 1) and appear to originate from weathered dolerite dikes. These flows are younger than Sirius Group sediments, which they truncate. At Table Mountain, moisture from windblown snow is < 5 cm per year. Surface temperatures are rarely above freezing and average between −25 and −30 °C (Pringle et al., 2003). Both the Sirius Group and the debris flows are ice-cemented 10–50 cm below the ground surface. High sublimation rates create the ice-free horizon near the surface, but in the subsurface, ice is being recharged via brine infiltration from snowmelt (Hagedorn et al., 2010).

2.1. Sampling

Cores were drilled (Dickinson et al., 1999) in November 2000 at two sites about 60 m apart with similar microclimates (Fig. 1). Core TM1 (77°57’36”S; 161°57’15”E, 1840 m above mean sea level) was drilled 4.6 m into ice-cemented debris flow sediments, which are very poorly sorted diamicts composed mostly of doleritic particles ranging in size from boulders to clay (Fig. 2).
Large (2–3 cm²) pockets of ice are visible in the frozen core, and water loss suggests a bulk porosity of 20–30% for these sediments (Supplementary Table 1).

Core TM4 (77°57′37″S; 161°57′20″E, 1845 m a.s.l) was drilled 3.2 m into ice-cemented, glaciofluvial sandstones of the Sirius Group, which are moderately-well sorted and coarse to medium grained with a few thin lenses of doleritic gravels (Fig. 2; Goff et al., 2002). Estimates from water loss indicate that porosity is 10–12% in these sediments (Supplementary Table 1). At each site, the dry permafrost was excavated and sampled by hand down to ice-cemented sediment (this was 0.44 m and 0.48 m below the surface for sites TM1 and TM4, respectively). Once extracted, the cores (diameter = 5.8 cm) were brushed clean, described, and cut (perpendicular to length), using a hydraulic splitter, into approximate 5 cm length and put into sealed plastic jars. Core segments were measured and weighed to determine density. Selected segments of core were placed in specially designed, plastic centrifuge jars, thawed, and spun (2000 rpm for 10 min) to extract pore water, which was previously frozen. Chemical analyses for anion abundance were run on this water (Supplementary Table 1).

After water extraction, the core segments were dried and re-weighed. These dry segments were friable and care was needed to extract vertically oriented samples, which were large (3 cm³) and intact. These samples were then vacuum impregnated with blue-dyed epoxy for thin sectioning.

### 2.2. Analyses

None of the core samples contained carbonates. Approximately 30 g of the dried core samples were gently crushed, and the large (> 2 cm) grains were removed. The samples were then washed repeatedly with boiling water to remove salts, after which they were freeze-dried. Because many samples were poorly sorted, they were sieved to exclude the > 2 mm fraction for grain size analysis (Supplementary Table 1). Washed solutions containing salt were evaporated and weighed to determine % salt. Some of these samples were reconstituted, filtered and measured for Be. The absence of detectable beryllium indicates that none was lost from the samples by washing them with boiling water.

The salt-free fraction was sieved at 0.5 mm. Be was leached from approximately 10 g of the < 0.5 mm fraction using hot 6 M HCl (Ditchburn and Graham, 2003). The difference in weight before and after this leaching is termed as the leached fraction (Supplementary Table 1). Concentrations of ⁹Be were determined by ICP-OES on a small aliquot of each leachate. ¹⁰Be concentrations were determined via accelerator mass spectrometry using the NIST standard SRM 4325 for calibration (Nishizumi et al., 2007; Zondervan et al., 2007).

### 3. Results and discussion

#### 3.1. Physical and chemical characteristics of the sediments

Sirius Group and Neogene sediments at Table Mountain have undergone several post-depositional alterations (Dickinson and Grapes, 1997), but the development of clay (fine particles < 10 µm) fabric is particularly relevant to the accumulation of meteoric ¹⁰Be. These fabrics are common throughout both cores, and microscopically, the clays are clearly illuviated. In Sirius Group sediments, clay laminae may be 10–20 µm thick, and connect grains by forming bridges several 100 µm long (Fig. 3A). Many laminae are highly birefringent. This is characteristic of illuviated clays, which are stacked or aligned on their flat sides to give unit extinction parallel to their axis (Fig. 3B). Clay drapes and bridges, which are shorter and thicker, are less common (Fig. 3C). In the debris flow sediments, clay coating and clumping is the dominant fabric (Fig. 3D). These coatings and clumps of clay show minor birefringence, which is not as bright as in Sirius Group laminae.

The translocation of clays to depths of > 4 m is difficult to explain because percolating waters from the surface do not penetrate more than 1–2 m in most soils, and if they do, the climate is so humid that clay translocation is uncommon. At Table Mountain, clay translocation cannot occur by cryoturbation because the ¹⁰Be and salt profiles are not homogenized and show distinct trends (Fig. 2). In the MDV with low erosion rates and millions of years of stability, deep clay translocation may result from low rates of seepage over long periods of time.

The two cores are compositionally, texturally, and genetically different, but their chemical profiles are remarkably similar (Fig. 2). Total anions in both cores peak at 1–1.5 m depth, and this probably results from the climatic conditions before the sediments became frozen. The profiles show that ⁹Be leaches from the sediment, whereas ¹⁰Be derives from the surface. Higher ¹⁰Be concentrations in TM1 than in TM4 may result from differences in both sediment type and erosion rates. ¹⁰Be concentrations in the subsurface are an order of magnitude lower than the surface, indicating the present lack of ¹⁰Be flux into the subsurface. Both profiles are essentially frozen in time, and ¹⁰Be in the subsurface has been left to decay since it was ‘closed’ off from the surface.

#### 3.2. ¹⁰Be erosion rates

Numerous variables that control accumulation of meteoric ¹⁰Be in soils limit its use as a dating tool (Graly et al., 2010). However, the flux of meteoric ¹⁰Be into the soil profile is one important variable that has essentially been eliminated since 3.9 Ma in the frozen environment of the Dry Valleys (Schiller et al., 2009). Close proximity (60 m) of the cored profiles allows us to make two important assumptions: (1) the flux (Φ) of meteoric ¹⁰Be at the surface is the same for each profile and (2) ¹⁰Be was...
sealed or frozen at the same time \((t)\) in each profile. These constraints, which are not possible using a single profile, allow calculation not only of a common age for closure but also of erosion rates since closure.

Over time, \(^{10}\text{Be}\) inventory of the soil surface reaches a steady state or equilibrium where \(^{10}\text{Be}\) flux \((Q)\) from the atmosphere is balanced by loss from (1) erosion, (2) transfer to the subsurface, and (3) radiogenic decay \((\text{Pavich et al., 1986})\). Assuming a relatively constant flux of \(^{10}\text{Be}\) and a minimal effect from decay, Willenbring and von Blanckenburg \((2010)\) described erosion rate \((E_r)\) as

\[
E_r = \frac{Q}{\rho N_{\text{modern}}}.
\]

Using soil densities \((\rho)\) at the surface and the \(^{10}\text{Be}\) flux \((Q)\) from Taylor Dome \((\text{Steig et al., 1995})\), modern \(^{10}\text{Be}\) concentrations at the surface \((N_{\text{modern}})\) indicate erosion rates of \(370 \pm 50 \text{ cm Myr}^{-1}\) at TM1 and \(1240 \pm 160 \text{ cm Myr}^{-1}\) at TM4 (Table 1).

This difference in calculated erosion rates relates directly to the differences in densities and surface concentrations of \(^{10}\text{Be}\) at each site. Sirius Group sediments at TM4 are friable and this may allow them to erode faster than the debris sediments at TM1, which are covered by desert pavement. This would also cause the surface concentration of \(^{10}\text{Be}\) at TM4 to be lower than at TM1.

\(^{10}\text{Be}\) erosion rates determined from Eq. \((1)\) are nearly two orders of magnitude higher than denudation rates for the MDV determined by Ivy-Ochs et al. \((1995)\) and Summerfield et al. \((1999)\) from in situ nuclides. This discrepancy, also observed by Graham et al. \((2002)\) and Schiller et al. \((2009)\), is best explained by the lack of \(^{10}\text{Be}\) transfer from the atmosphere to the subsurface because arid, windy conditions allow little \(^{10}\text{Be}\) aerosol to accumulate on the soil surface. Hence, \(^{10}\text{Be}\) erosion rates determined from the known meteoric flux of Taylor Dome, which is an accumulation area, do not represent surface denudation in the MDV. In other words, atmospheric flux \((Q)\) onto the surface, which is the same at each site, does not represent \(^{10}\text{Be}\) flux into the soil profile.

### 3.3. Closure ages

Closure ages may be calculated if (1) the climatic change or closure of \(^{10}\text{Be}\) is faster than \(^{10}\text{Be}\) equilibration at the surface; (2) the subsurface \(^{10}\text{Be}\) profiles have decayed without further input or modification, (3) the modern-surface \(^{10}\text{Be}\) concentration \((N_{\text{modern}})\) remains in a steady state at the surface, and (4) the difference between \(N_{\text{modern}}\) and the paleo-surface concentration of \(^{10}\text{Be}\) \((N_{\text{paleo}})\) results only from decay. In the MDV, these assumptions are reasonable, and time \((t)\) since closure can be determined by the relationship:

\[
t = -\ln(N_{\text{paleo}}/N_{\text{modern}})/\lambda_{10\text{Be}}\quad (2)
\]

where \(\lambda_{10\text{Be}}\) is the decay constant of \(^{10}\text{Be}\).

Projecting the \(^{10}\text{Be}\) concentrations in the frozen subsurface to the surface gives a first estimate for \(N_{\text{paleo}}\) (Fig. 4):

\[
N_{\text{paleo}} = ae^{bt}\quad (3)
\]

where constants \((a)\) and \((b)\) are found from fitting this equation to the measured \(^{10}\text{Be}\) concentrations in each subsurface profile (Table 1). Erosion \((E)\) is a length measurement without the time component found in Eq. \((1)\). In the regression equations (Fig. 4), note that erosion \((E > 0)\) is exchangeable with depth \((E < 0)\). Because \(^{10}\text{Be}\) concentrations decrease exponentially with depth,

<table>
<thead>
<tr>
<th>Function</th>
<th>TM 1</th>
<th>TM 4</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Erosion rate</td>
<td>(1.30 \times 10^5)</td>
<td>(2.0)</td>
<td>(\text{atom cm}^{-2} \text{a}^{-1})</td>
</tr>
<tr>
<td>(^{10}\text{Be}) flux ((Q))</td>
<td>(1.2)</td>
<td>(5.23 \pm 0.10 \times 10^7)</td>
<td>(\text{g cm}^{-3})</td>
</tr>
<tr>
<td>Soil density ((\rho))</td>
<td>(2.90 \pm 0.10 \times 10^8)</td>
<td>(1240 \pm 160)</td>
<td>(\text{atom g}^{-1})</td>
</tr>
<tr>
<td>(N_{\text{modern}})</td>
<td>(370 \pm 50)</td>
<td>(1.66 \pm 0.14 \times 10^5)</td>
<td>(\text{cm Myr}^{-1})</td>
</tr>
<tr>
<td>Erosion rate ((E))</td>
<td>(6.1 \pm 0.5)</td>
<td>(6.8 \pm 0.2)</td>
<td>(\text{Myr})</td>
</tr>
<tr>
<td>Closure age ((t))</td>
<td>(1.31 \pm 0.27 \times 10^7)</td>
<td>(1.66 \pm 0.14 \times 10^5)</td>
<td>(\text{atom g}^{-1})</td>
</tr>
<tr>
<td>Minimum closure age ((t))</td>
<td>(11.0 \pm 8.6)</td>
<td>(6.2 \pm 5.1)</td>
<td>(\text{cm Myr}^{-1})</td>
</tr>
<tr>
<td>Vertical erosion ((E))</td>
<td>(1.9 \pm 1.6)</td>
<td>(5.9 \pm 5.6)</td>
<td>(\text{Myr})</td>
</tr>
<tr>
<td>Closure age ((E))</td>
<td>(4.3 \pm 0.44 \times 10^6)</td>
<td>(4.2 \pm 3.4)</td>
<td>(\text{atom cm}^{-2} \text{a}^{-1})</td>
</tr>
<tr>
<td>Maximum closure age ((t))</td>
<td>(1.3 \pm 1.1)</td>
<td>(4.2 \pm 3.4)</td>
<td>(\text{atom cm}^{-2} \text{a}^{-1})</td>
</tr>
<tr>
<td>Vertical erosion ((E))</td>
<td>(1.8 \pm 1.5 \times 10^3)</td>
<td>(8.6 \pm 6.6)</td>
<td>(\text{Myr})</td>
</tr>
<tr>
<td>Influence of in situ produced (^{10}\text{Be})</td>
<td>(4)</td>
<td>(4)</td>
<td>(\text{atom cm}^{-2} \text{a}^{-1})</td>
</tr>
<tr>
<td>Regression constants ((a))</td>
<td>(8.65 \pm 4.05 \times 10^6)</td>
<td>(3.39 \pm 0.51 \times 10^7)</td>
<td>(\text{cm Myr}^{-1})</td>
</tr>
<tr>
<td></td>
<td>(1.7 \pm 1.7 \times 10^{-2})</td>
<td>(1.4 \pm 1.4 \times 10^{-2})</td>
<td>(\text{atom cm}^{-2} \text{a}^{-1})</td>
</tr>
<tr>
<td></td>
<td>(2.83 \pm 0.11 \times 10^8)</td>
<td>(5.08 \pm 0.11 \times 10^7)</td>
<td>(\text{cm Myr}^{-1})</td>
</tr>
<tr>
<td>Vertical erosion ((E))</td>
<td>(\sim 3.4 \times 10^8)</td>
<td>(\sim 171)</td>
<td>(\text{cm Myr}^{-1})</td>
</tr>
<tr>
<td>Erosion rate ((E))</td>
<td>(\sim 51)</td>
<td>(\sim 33)</td>
<td>(\text{atom cm}^{-2} \text{a}^{-1})</td>
</tr>
<tr>
<td>Closure age ((E))</td>
<td>(\sim 5.1)</td>
<td>(\sim 5.1)</td>
<td>(\text{Myr})</td>
</tr>
</tbody>
</table>

Errors are 2 std. dev. of the mean.

\(^{5}\text{Steig et al. (1995).}\)
an exponential fit gave excellent $R^2$ values (Fig. 4). The degree of exponential decrease with depth is represented by the fit parameter $b$ in Eq. (3) and represents the combined effect of radiometric decay and downward transport of $^{10}\text{Be}$ (Willenbring and von Blanckenburg, 2010).

Substituting measured $N_{\text{modern}}$ and projected $N_{\text{paleo}}$ (at $E=0$) into Eq. (2) gives closure ages of $6.1 \pm 0.5$ Ma for TM1 and $6.8 \pm 0.2$ Ma for TM4 (Table 1). The proximity of the profiles suggests that these closure ages should be identical, yet they are barely within error of each other. This discrepancy results from ignoring the fact that these cores have eroded differently since closure. We can correct for erosion by assuming that the ratio of surface erosion rates at these two sites did not vary with time since close-off and that both profiles closed off at the same time.

By assuming that the closure age ($t$) is the same for each profile, we can rewrite Eq. (2) as

$$N_{\text{paleo}1}/N_{\text{modern}1} = N_{\text{paleo}4}/N_{\text{modern}4},$$

where subscripts 1 and 4 refer to quantities from profiles TM1 and TM4, respectively. In Eq. (4), the unknown, $N_{\text{paleo}}$ must be adjusted so that the ratio $N_{\text{paleo}}/N_{\text{modern}}$ is the same for each profile when $t$ is common. If atmospheric $^{10}\text{Be}$ flux ($Q$) is the same for each profile, then the ratio of erosion rates (Eq. (1)) of the two profiles is inversely proportional to $N_{\text{modern}}$ and soil density in each profile

$$E_1/E_4 = Q/\rho_1 N_{\text{modern}1}/Q/\rho_4 N_{\text{modern}4} = \rho_4 N_{\text{modern}4}/\rho_1 N_{\text{modern}1}$$

(5)

Importantly, when $Q$ is constant, erosion rates may change through time, but they must change to the same degree so that their ratio remains constant. The amount of erosion in one profile with respect to erosion in the other profile is now

$$E_4 = \rho_4 N_{\text{modern}4}/\rho_1 N_{\text{modern}1} E_1.$$ 

(6)

Elimination of $N_{\text{paleo}1}$, $N_{\text{paleo}4}$, and $E_1$, from Eqs. (3), (5) and (6) yields

$$E_1 = \frac{\ln(a_1 N_{\text{modern}4}/a_4 N_{\text{modern}1})}{b_1 - b_2 (\rho_4 N_{\text{modern}4}/\rho_1 N_{\text{modern}1})},$$

(7)

Substitution of $E_1$ into Eq. (3) and using (a) and (b) determined for TM1 gives $N_{\text{paleo}1}$. Substitution of $N_{\text{paleo}1}$ into Eq. (2) then gives the common value of time ($t$). Adjusted erosion rates ($E_r$) are determined from the common closure age ($t$) and the total amount of erosion ($E$) since closure for each profile (Table 1):

$$E_r = E/t$$

(8)

The common closure age determined from this erosion-corrected $N_{\text{paleo}}$ is $5.9 \pm 0.6$ Ma. This represents a minimum age because it assumes that $N_{\text{modern}}$, which is a low concentration, has not changed since closure. However, more $^{10}\text{Be}$, which would increase closure age, may have been retained at the soil surface had the climate been wetter in the past. An upper age limit is determined by increasing $N_{\text{modern}}$ for both profiles by a factor of four. This brings $^{10}\text{Be}$ surface concentrations to the maximum as found in temperate soils (Graly et al., 2010; Pavich et al., 1986).

Substituting these maximum values into Eq. (2) and keeping the total erosion constant increases the common closure age to $8.6 \pm 0.6$ Ma. While the present-day $N_{\text{modern}}$ is known, the past $N_{\text{modern}}$, which was the surface concentration at the time of closure, is the main unknown in our model. Bracketing $N_{\text{modern}}$ between high and low values constrains the closure age between roughly 6 and 9 Ma.

The common closure ages are determined above yield erosion rates of 1–2 cm Myr$^{-1}$ for TM1 and 4–6 cm Myr$^{-1}$ for TM4 (Table 1). These erosion rates are slightly lower than those reported by Ivy-Ochs et al. (1995) and Summerfield et al. (1999). Substituting our erosion rates into Eq. (1) gives a $^{10}\text{Be}$ flux into the soil ($Q_e$) of $6.6 \times 10^9$ atom cm$^{-2}$ a$^{-1}$, which is about 0.5% of the measured meteoric flux at Taylor Dome and in good agreement with other such estimates (Graham et al., 2002; Schiller et al., 2009) in the MDV.

3.4. Influence of in situ produced $^{10}\text{Be}$

At Table Mountain where erosion rates are low and meteoric $^{10}\text{Be}$ has decayed for at least 6 Myr, an unknown amount of in situ $^{10}\text{Be}$ may have been extracted in our samples. This is because the strong acid leach used to remove meteoric $^{10}\text{Be}$ may also have removed an unknown amount of in situ $^{10}\text{Be}$ from within grain lattices due to
partial dissolution. Experiments to determine concentrations of meteoric and in situ $^{10}$Be in our samples are beyond the scope of this paper. Instead, we estimate the in situ component by (1) using the CRONUS calculator (Balco et al., 2008; CRONUS, Earth Project) for the production rate in quartz at Table Mountain, and by (2) assuming that this value applies equally well for the leached portion of our soil samples (Fig. 5; Supplementary Table 3).

Modeled in situ $^{10}$Be (atom/g of quartz) production with no erosion approximates was measured in the samples for TM4. On the other hand, the measured profile for TM1 is a factor of 2–3 above the in situ model. Moreover, explaining the observed depth dependence in TM1 by in situ production alone would require an unrealistically large decrease in fast-neutron attenuation length from 150 g cm$^{-2}$ to 100 g cm$^{-2}$ (Brook et al., 1996; Brown et al., 1992). Because different production parameters are needed in adjacent profiles with the same exposure, we conclude that in situ $^{10}$Be cannot be the main source of $^{10}$Be measured in the profiles.

By subtracting or removing modeled in situ $^{10}$Be from measured $^{10}$Be concentrations, closure ages can be re-calculated from the residual concentrations. For these calculations, only the upper three subsurface points of each profile are used for the regressions, where there is enough $^{10}$Be remaining after the correction (Fig. 5). Because the correction is relatively large, points below these upper three are significantly affected by their original uncertainties and poor resolution of the in situ corrections. Using these regressions and corrected surface concentrations, the re-calculated, minimum closure age for the profiles is about 5.1 Ma (Table 1). While this common age is similar to that calculated for meteoric $^{10}$Be, the resulting erosion rates and the flux (Q) are significantly different (Table 1).

Several aspects of the closure age method are also revealed by the in situ model. Different densities of the profiles (i.e., $c_1=1.2$ g cm$^{-3}$; $c_2=2.0$ g cm$^{-3}$) cause slightly different slopes of the in situ $^{10}$Be profiles, which are sub-parallel on a log scale. This means that the in situ $^{10}$Be correction is comparable to a baseline subtraction, where both profiles are similarly modified. The age calculation is based on the ratio of surface to subsurface concentrations in each profile. The age shift by including the in situ production estimate is minor because the baseline subtraction preserves the relative difference in the ratios. However, the erosion rate and flux are based on the relative difference between surface and subsurface concentrations within each profile. Thus, erosion and flux must increase significantly to accommodate the relative differences between surface and extrapolated paleo-surface concentrations.

Unknown amounts of in situ $^{10}$Be in our samples appear to have little effect on the age interpretation of our data. It is important to note that the recalculation of the closure age is based on a maximum contribution from in situ $^{10}$Be and that the true contribution might be much less. For situations where the meteoric $^{10}$Be concentrations are low, the leach method of Ditchburn and Graham (2003), used in this study, is preferred over the whole-sediment extraction method of Stone (1998). Some combinations of these methods may allow discrimination between meteoric and in situ $^{10}$Be, which has the potential to enhance confidence in modeling soil profiles.

4. Conclusions

Confidence in the modeled closure ages and erosion rates comes largely from the proximity of the genetically different soil profiles that have similar chemical and isotopic signatures. $^{10}$Be and illuvial clays, frozen in the modern profile, reflect a time prior to 6–9 Ma when liquid transport of salts and fine particles was possible in a warmer than present climate. Evidence from ANDRILL (McKay et al., 2009) paints a complex history of Late Miocene cooling at low elevation in the Ross Embayment, which is supported by similar evidence from other parts of Antarctica (Hambrey and Mckelvey, 2000). Although high elevation glacial stratigraphy provides compelling evidence of pronounced cooling at ca. 12 Ma (Lewis et al., 2007), variability in MDV paleoclimate since this time is not easily resolved. Our independent evidence confirms persistent polar conditions in high elevations of the MDV since Pliocene time, and implies an extreme climate contrast between the Transantarctic Mountains and the Ross Sea coast.

The advantage of using two profiles to understand meteoric $^{10}$Be is that the atmospheric flux of $^{10}$Be is essentially the same for each site. This allows comparison of other variables such as erosion and $^{10}$Be transfer into the subsurface. However, the value of the two-profile method needs testing in temperate climates where, unlike the MDV, $^{10}$Be currently transfers into the subsurface. In the MDV, inventories of meteoric $^{10}$Be determined from the modern flux at Taylor Dome are clearly inaccurate. In addition, where erosion is slow and exposure times are long, the production of in situ $^{10}$Be may increase $^{10}$Be concentrations by an unknown amount. Our in situ model shows that for the age calculation, which is based on relative differences between surface and subsurface $^{10}$Be concentrations in two adjacent profiles, in situ production has little impact. However, in situ production may influence the calculated erosion and flux of $^{10}$Be into the profile, which are based on each individual profile.
Overprint in situ production requires further investigation before results from the two-profile method can be fully accepted.

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Appendix A. Supporting information

Supplementary data associated with this article can be found in the online version at http://dx.doi.org/10.1016/j.epsl.2012.09.003.

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