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Using the sunspot cycle to date ice cores

Eric J. Steig

Institute of Arctic and Alpine Research, University of Colorado, Boulder, Colorado

David L. Morse[1] and Edwin D. Waddington

Geophysics Program, University of Washington, Seattle, Washington

Pratigya J. Polissar[2]

School of Natural Sciences, Hampshire College, Amherst, Massachusetts

Abstract. For many ice cores, the snow accumulation rate is too low to preserve annual stratigraphy, precluding direct measurement of annual layer thickness. For selected Holocene sections of the Taylor Dome core, East Antarctica, we instead use high-resolution ^{10}Be measurements to establish a nominal 11-year thickness, taking advantage of the solar cycle in cosmogenic isotope production. We compare measured thicknesses with the layer thickness profile predicted by a finite element ice flow model. The results are in good agreement, supporting the assumption that the length of the solar cycle has remained essentially constant throughout the Holocene. For ice cores where annual layers are not preserved, the ^{10}Be 11-year layer method can be used as an independent check on flow-model estimates of layer thickness and to estimate past accumulation rates. It should be possible to accurately date ice cores by counting 11-year layers detected with continuous high-resolution ^{10}Be measurements.

Introduction

Following burial and firnification at the surface of a glacier or ice sheet, annually accumulated layers of snow undergo thinning due to glacier flow. Each glacier or ice sheet has a characteristic time scale, T, for ice deformation given by

$$T = \frac{H}{\dot{b}} \qquad (1)$$

where H is the ice thickness and $\dot{b}$ is the ice-equivalent accumulation rate. Typical values of T range from a few hundred years for temperate valley glaciers to ~10^5 years for the large polar ice sheets. The thinning process can be described by a function that gives the ratio between the thickness at depth and the initial (ice equivalent) thickness at the surface. The thinning function is usually estimated with a geophysical ice flow model which calculates the total vertical strain experienced by ice particles as they travel downward from the surface [*Waddington et al.*, 1993]. For ages less than T, layer thicknesses can generally be predicted with confidence.

Ice cores are drilled to meet two possible goals: recovering high-resolution records and recovering old ice. High-resolution records require sites where the accumulation rate, $\dot{b}$, is relatively high (> 0.2 m ice a^{-1}). Under these conditions, annual stratigraphic layers are generally preserved in the ice, and their thickness can be measured directly using seasonal indicators such as δ^{18}O cycles [*Hammer et al.*, 1986] or visible differences between summer and winter snowfall [*Alley et al.*, 1993]. The age-depth relationship for the core can then be obtained by direct layer counting. Also, differences between observed and predicted layer thickness can be used to determine accumulation rate history [*Cutler et al.*, 1995], essential for estimating fluxes of chemical tracers from their observed concentrations [*Alley et al.*, 1995; *Waddington*, 1996]

Often, the goal of obtaining high-resolution records conflicts with the goal of recovering old ice. An example is the Taylor Dome ice core, retrieved in 1994 from a local ice-accumulation area near McMurdo Sound, Antarctica (77°59'S, 158°59'E). This core is of potentially great importance because of its proximity to the Ross Sea embayment, where dramatic changes in ocean and ice-sheet conditions have taken place through glacial-interglacial cycles [*Denton et al.*, 1989; *Steig et al.*, 1997]. At Taylor Dome (T ~ 10^4 years) ice flow models can reconstruct Holocene ice flow patterns with reasonable precision. However, the accumulation rate ($\dot{b}$ = 0.07 m ice a^{-1}) is too low to preserve annual stratigraphy beyond the upper 20 meters of firn [*Grootes and Steig*, 1992].

In this paper, we present a new method for measuring layer thicknesses in low-accumulation ice cores. Using Taylor Dome as a test case, we determine an 11-year stratigraphy using high-resolution measurements of cosmogenic ^{10}Be. We demonstrate the applicability of the method by comparing measured layer thicknesses with results from a finite element ice flow model.

The 11-year cycle

The essence of our approach is the recognition that the production of ^{10}Be by cosmic-ray interactions with atmospheric N, O and other elements is periodic [*Lal and Peters*, 1967], and that this periodicity can be detected in ^{10}Be deposited at the earth's surface. Of particular interest is the well-known Schwabe sunspot cycle, having a period averaging 11 years. Averaged over the entire earth, the amplitude of ^{10}Be production over the course of a typical 11-year cycle is about 10%. At polar latitudes, this amplitude increases to > 20 % and is therefore relatively easy to detect in polar ice cores. Other changes in production rate are

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[1] Now at Institute for Geophysics, University of Texas, Austin, Texas.

[2] Now at Department of Geosciences, University of Massachusetts, Amherst, Massachusetts.

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either much smaller in amplitude (the 90 year Gleissberg solar cycle [*Beer et al.*, 1994]), are non-periodic (short-lived periods of anomalously high production [*Raisbeck et al.*, 1987]), or occur over much longer time scales (changes in geomagnetic field strength [*Raisbeck and Yiou*, 1988]). Historical records of sunspot and aurora observations show that the 11-year cycle has persisted for at least the last several hundred years [*Eddy*, 1988]. The 11-year cycle has been detected in ^{10}Be measurements in ice cores from both Greenland [*Beer et al.*, 1990] and Antarctica [*Steig et al.*, 1996]. *Attolini et al.* [1988] have shown that the 11-year cycle in ^{10}Be deposition persisted through the Maunder sunspot minimum, when the visible sunspot cycle had disappeared. Our working hypothesis is that the historical behavior of the 11-year sunspot cycle has persisted through the characteristic timescale of the Taylor Dome core, about 10^4 years.

Determination of layer thickness with ^{10}Be

Section *A* in Figure 1 shows the results of ^{10}Be analyses from a shallow firn core at Taylor Dome. The periodicity in section *A* has been independently dated using seasonal δ^{18}O stratigraphy and identification of AD 1954 and 1964 bomb-radioactivity horizons, and is demonstrably associated with the 11-year solar cycle [*Steig et al.*, 1996]. Additional ^{10}Be analyses were conducted at 49.12-51.25 m, 130.00-131.00 m, 220.12-221.12 m and 298.60-299.80 meters in the Taylor Dome ice core. We refer to these sections as *B* through *E* in order of increasing depth. Samples were processed and analyzed by accelerator mass spectrometry (AMS) according to the methods described by *Steig* [1996], following *Raisbeck et al.* [1978] and *Southon et al.* [1990]. Precision of the analyses is better than ±5 %. Section *B* was sampled at a resolution of 10 m^{-1}; sections *C*, *D* and *E* were sampled at 20 m^{-1}. The Pleistocene-Holocene transition in the Taylor Dome ice core occurs at a depth of ~375 m, as readily observed in ice fabric [*Fitzpatrick*, 1994] and geochemical profiles [*Steig*, 1996; *Mayewski et al*, 1996]. Thus, the sections analyzed are all from the Holocene period.

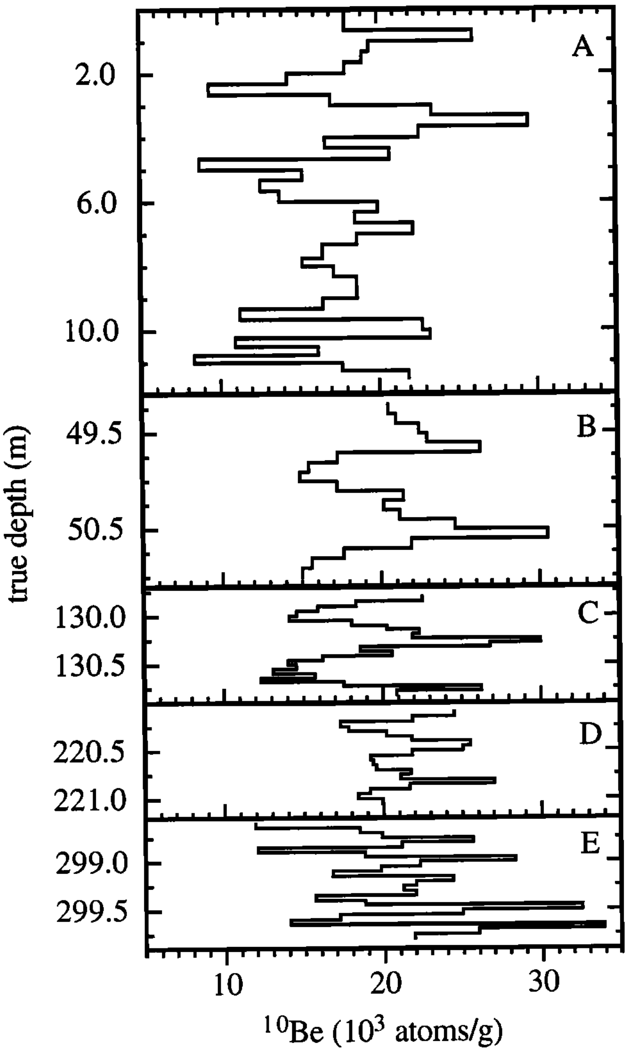

Figure 1. ^{10}Be concentrations for selected depth intervals at Taylor Dome. The panel labels "A", "B", etc. correspond to the section names referred to in the text.

Figure 1 shows that there is a clear decrease in the wavelength of ^{10}Be variations with depth. Neither high-resolution δ^{18}O [*Steig et al.*, 1996] nor electrical conductivity measurements [*K. Taylor*, personal communication, 1995] on the Taylor Dome core show evidence of periodicity at or near that observed in ^{10}Be. We conclude that the obvious, strong cycles in sections *B* through *E*, having amplitudes comparable to those in section *A*, cannot be explained by meteorological factors, but likely reflect Schwabe-cycle production-rate variations. The observed amplitudes are as large as, or larger than, predicted for production of ^{10}Be in the polar atmosphere. The significance of this result with respect to atmospheric mixing processes is discussed by *Steig et al.* [1996].

We determine the average wavelengths, $\overline{\Delta z}$, of ^{10}Be variations in sections *B* through *E* by measuring peak-to-peak and trough-to-trough distances. In the event that a peak or trough is defined by more than one data point, the average depth is adopted. We use measured densities, ρ [*Fitzpatrick*, 1994], to calculate the ice-equivalent 11-year layer thickness, λ_{11}, as a function of ice equivalent depth, z^*. Results are shown in Table 1 and plotted in Figure 2. Error bars take into account both analytical uncertainty and the variable length of the nominal 11-year cycle. From the historical sunspot data [*Stephenson*, 1990], we take this cycle length as $11 \pm 2/\sqrt{n}$ years (1 standard deviation) for n consecutive cycles. We emphasize that we are assuming, as a working hypothesis, that the historical behavior of the sunspot cycle is representative of the pre-historical period. The possibility of larger variations in cycle length could be tested with ^{10}Be measurements in annual-layer-counted cores from central Greenland.

Determination of layer thickness with an ice flow model

We use the two-dimensional finite element flow model developed by *Raymond* [1983] to model the flow of ice passing through the Taylor Dome core site [*Waddington et al.*, 1993]. The model uses a Glen-type flow law [e.g. *Paterson*, 1994]:

$$\dot{\varepsilon} = A\tau^n \qquad (2)$$

where $\dot{\varepsilon}$ is the effective strain rate, A is the temperature-dependent rheological softness parameter, τ is the effective shear stress and the stress exponent, n, is equal to 3. The model geometry is provided by detailed surface and bedrock topography from ground-based surveys and ice-penetrating radar [*Morse and Waddington*, 1992; *Morse*, 1997]. The model allows the surface to evolve to steady-state with the modern accumulation rate pattern, taken from a network of accumulation stakes, shallow cores and snow pits [*Grootes and Steig*, 1992; *Waddington and Morse*, 1994; *Morse*, 1997]. *A* is adjusted until the steady-state surface topography matches the modern profile. We calculate the tensor

Table 1. Measured 11-year layer thickness at Taylor Dome. Uncertainties (± 1 standard deviation) take into account measurement precision and the variable length of the 11-year cycle.

Section Name	True Midpoint Depth, z (m)	Ice Equivalent Depth, z^* (m)	11-year Layer Thickness, $\Delta\bar{z}$ (m)	Density, ρ (kg m^{-3})	Ice Equiv. 11-year Thickness, λ_{11} (m)
A	6.00	2.7	2.04 ± 0.20	354.30 ± 5	0.79 ± 0.10
B	49.50	32.4	0.93 ± 0.05	712.75 ± 5	0.72 ± 0.10
C	130.50	108.0	0.55 ± 0.02	908.78 ± 5	0.54 ± 0.07
D	220.50	197.4	0.39 ± 0.02	914.45 ± 5	0.39 ± 0.05
E	299.55	276.0	0.24 ± 0.04	916.95 ± 5	0.24 ± 0.04

finite strain experienced by ice particles along trajectories that intersect the ice core location, and use the layer normal strain to determine the layer thinning function.

We use the layer thinning function and an assumed constant accumulation rate of b = 0.07 m ice a^{-1} (the measured modern value at the core site) to calculate the 11-year layer thickness profile. We emphasize that the difference between the true and assumed accumulation rate—not the shape of the model-derived thinning function—is the chief uncertainty in this calculation. The assumption that b is constant is reasonable. For time intervals much longer than 11 years, average b is the dominant controlling factor on average ^{10}Be concentration [*Raisbeck and Yiou*, 1985]. For ~200 year averages, Holocene ^{10}Be concentrations in the Taylor Dome core vary by less than 10 % [*Steig*, 1997].

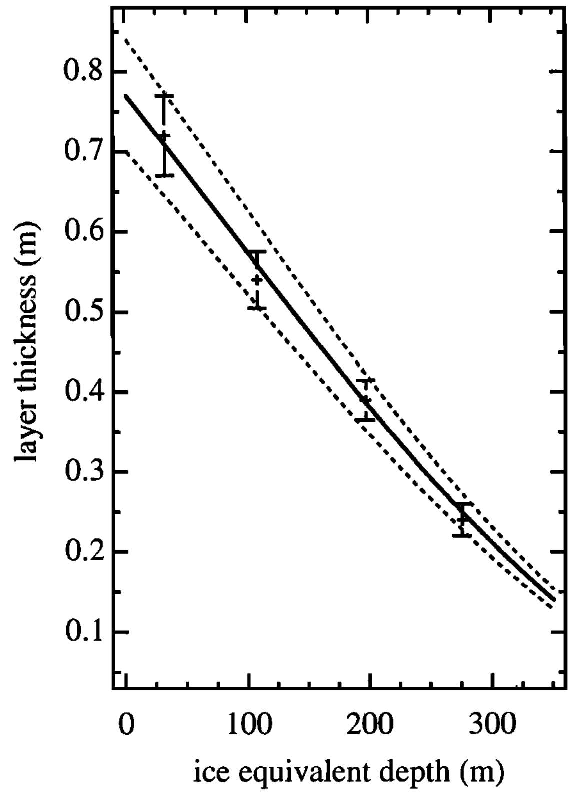

Figure 2. Comparison of 11-year layer thicknesses (λ_{11}) from high-resolution ^{10}Be data (+) and calculated layer thickness profile (Λ_{11}) from a finite element ice flow model (——) assuming a constant Holocene accumulation rate of 0.07 m ice equiv. per year. Error bars show uncertainty (0.5 standard deviations) of the measured thicknesses, taking into account the variable length of the nominal 11-year cycle. Upper and lower dashed lines show calculated 12 and 10-year layer thickness profiles, respectively.

Among the five shorter time intervals where we have measured ^{10}Be at higher resolution to determine the nominal 11-year layer thickness, the average concentration varies by less than 5 %. We plot the derived layer thickness profile, Λ_{11}, as a function of z^* in Figure 2. For comparison, we also show calculated 10 and 12-year layer profiles in Figure 2, again using $\dot{b}$ = 0.07 m ice a^{-1}. Clearly, it is the 11-year layer thickness profile that corresponds most closely to the measured ^{10}Be wavelengths.

Discussion

The agreement of the measured ^{10}Be wavelengths and the predicted 11-year layer thickness is excellent. As Figure 2 illustrates, the profile from the glacier flow model fits the data, in all cases, to within less than 0.5 standard deviations. This comparisons provide a compatibility test of both the assumptions of the flow model and the uniformity of the Schwabe period through the Holocene. We conclude the following: (1) The cyclicity of ^{10}Be concentrations measured in the Taylor Dome core is due to Schwabe-cycle (11-year) production rate variations. (2) The assumption of constant accumulation for the time intervals considered, the shape of the model-derived thinning function, and the assumed stability of the Schwabe period are either mutually correct or are deviously in error such that their effects cancel to give the impression of the former.

Valuable absolute chronologies and accumulation histories derived from annual layer counting and layer thickness observations have been obtained for ice cores from high accumulation areas in central Greenland. While existing and planned Antarctic ice cores can provide a wealth of paleoclimate information over time scales of T ~ 10^4 to 10^5 years, annual stratigraphy is generally not preserved at sites with these desirable long times scales. In principle, a depth-age scale based on counted 11-year solar cycles could be produced for such sites by continuous high-resolution ^{10}Be sampling. This may currently be impractical due to the high cost of ^{10}Be analyses. In the meantime, high-resolution ^{10}Be sampling at judiciously spaced depth intervals tightly constrains the depth-age curve by measuring its slope and, when used with thinning functions from ice flow models, provides a means to determine past accumulation rates.

Continuous ^{10}Be sampling may be the only technique capable of providing a layer-counted chronology for important climate events such as the "Antarctic Cold Reversal" [*Jouzel et al.*, 1995] in low-accumulation cores from East Antarctica. As an example, at a depth of 380 meters in the Taylor Dome ice core, the age is about 15,000 years [*Steig*, 1996]. At this depth, the annual layer thickness is about 0.2 cm and the ^{10}Be concentration is about 10^5 atoms per gram of ice. Given that ½ the cross sectional area of the 15 cm diameter core is available, sampling at single-year resolution yields 15 g of ice, or about 1.5 × 10^6 atoms of ^{10}Be, measurable at current AMS detection limits. Since it may be reasonable to define 11-year cycles on the basis of biannual resolution (i.e. 5.5 samples per cycle), it should be possible to extend the limit of our method to about 30,000 years at Taylor Dome. At other sites in East Antarctica, accumulation rates are often lower, but ^{10}Be concentrations are correspondingly higher. This, and the fact that thinning is less extreme where the total ice thickness is greater, means that the practical limit of our method at sites such as Vostok may be as great as ~50,000 years.

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E. J. Steig, Institute of Arctic and Alpine Research, University of Colorado, Boulder, CO 80309. (e-mail: steig@stripe.colorado.edu)

D. L. Morse, Institute for Geophysics, University of Texas, 4412 Spicewood Springs Rd., Bldg. 600, Austin, TX 78759. (e-mail: morse@ig.utexas.edu)

E. D. Waddington, Geophysics Program 351650, University of Washington, Seattle, WA 98185 (e-mail: edw@geophys.washington.edu)

P. J. Polissar, Department of Geosciences, University of Massachusetts, 233 Morrill Science Center, Amherst, MA 01003. (e-mail: polissar@geo.umass.edu)

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