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SURFACE EXPOSURE DATING WITH COSMOGENIC NUCLIDES

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Abstract

Recent developments in accelerator mass spectrometry and noble gas mass spectrometry now allow precise analysis of cosmic-ray produced nuclides in terrestrial materials. Measurement of ^{3}He, ^{21}Ne, ^{10}Be, ^{26}Al, ^{14}C, and ^{41}Ca can be used to determine exposure ages, erosion rates, and burial histories of geological surfaces in favorable situations. The ability to measure surface exposure ages is of particular importance to many aspects of geology, geomorphology, and archeology. In order to calculate the exposure age for a terrestrial surface, the production rates must be well known, the sample must be a closed system for the isotope of interest, and the sample must have had a simple exposure to cosmic rays (no burial or cover). The production rates of all cosmogenic nuclides vary as a function of altitude, latitude, and time, and are critical to all applications; better calibrations of the production rates, by measurements of surfaces of known age, will be necessary to improve the accuracy and precision of surface exposure dating. No single cosmogenic nuclide is applicable to all dating situations; each has its own limitations, related to age of the materials, the half-life of the nuclide, the mineral phases present, and the difficulty of the measurements. Existing data also demonstrate that measurement of more than one cosmogenic nuclide, in the same samples, is extremely useful in evaluating the assumptions inherent to the technique. A number of recent studies demonstrate that surface exposure dating can provide age information for surfaces as young as several thousand years and as old as millions of years. Although many of the successful applications have been within the fields of glacial geology and geomorphology, the technique should also be extremely useful to other fields, such as archaeology. This chapter outlines the assumptions and uncertainties inherent to surface exposure dating with cosmogenic nuclides.

Introduction

ALTHOUGH USING COSMOGENIC NUCLIDES AS A means of dating terrestrial surfaces is not a new idea (see Davis and Schaeffer 1956), the method has received considerable attention recently. This is in large part due to the development of accelerator mass spectrometry (AMS) for the measurement of small amounts of cosmogenic radionuclides (for example, ^{10}Be, ^{26}Al, ^{36}Cl, ^{14}C), making it possible to apply this technique to the low levels found in terrestrial surfaces (see Elmore and Phillips 1987; Litherland 1987; Raisbeck et al. 1987). Prior to this development, dating using cosmogenic nuclides had been applied primarily to studies of meteorites and lunar samples, which are exposed to much higher cosmic-ray fluxes than exist at the earth's surface, and consequently have higher concentrations of cosmogenic nuclides (Eugster 1988; Honda and Arnold 1964; Mazor et al. 1970; Reedy et al. 1983; Vogt et al. 1990). In addition to the radiogenic nuclides mentioned above, two stable cosmogenic noble gas isotopes, ^{3}He and ^{21}Ne, have recently been detected in terrestrial surfaces using conventional noble gas mass spectrometers, instruments that are considerably less expensive than accelerator mass spectrometers (Craig and Poreda 1986; Graf et al. 1991; Kurz 1986a, 1986b; Marti and Craig 1987).

As a result of these technical advances, there are now a number of different cosmogenic nuclides that can be used for dating surfaces, each with its own set of advantages and disadvantages. These techniques have been applied primarily to geological surfaces, such as glacial moraines, striated bedrock, and lava flows, but could also be successfully applied to archeological artifacts and stone buildings. The purpose of this chapter is to review the various methods of using the different nuclides, to summarize the uncertainties that are common to all the methods, and to give several examples that illustrate both the potential and the limitations of the techniques. The emphasis here is on the use of ^{3}He, ^{10}Be, and ^{26}Al because these are the nuclides for which there is the most data, apart from ^{36}Cl which is discussed by Zreda and Phillips (chapter 8 of this volume). New methods based on *in situ* ^{14}C and ^{41}Ca are in development (Fink et al. 1990; Jull et al. 1991) and are discussed briefly.

Principles of Surface Exposure Dating with Cosmogenic Nuclides

The basic principle of surface exposure dating with cosmogenic nuclides is simple. Assuming that the cosmic-ray flux is constant with time, the concentration of accumulated cosmogenic nuclides within surficial rocks is directly related to the time the surface has been exposed. For the purpose of this article, the term "cosmogenic" is intended to refer to any nuclide that is produced by cosmic-ray particles. There are a number of different nuclear reactions that are important in the earth's atmosphere and at the earth's surface, including neutron capture, muon capture, and spallation. The dominant production mechanism within rocks in the top two meters of the earth is neutron-induced spallation, which is simply a collision of a cosmic-ray neutron with a target nucleus, resulting in the breaking apart of the nucleus into fragments. This process is illustrated graphically in Figure 7.1, which shows a spallation event in a photographic emulsion flown at high altitude. The energy threshold for spallation is typically 20 to 50 Mev, and the spallation cross sections are not strongly composition-dependent (because spallation is essentially a collision process). A list of the cosmogenic nuclides that have been most commonly used for surface exposure dating is given in Table 7.1. With the exception of ^{36}Cl, which has a significant production by neutron capture in some rocks (on ^{35}Cl, see Zreda and Phillips, chapter 8 of this volume), and ^{41}Ca, produced dominantly by thermal neutron capture in ^{40}Ca (Fink et al. 1990), spallation is the most important production mechanism for these nuclides. As discussed below there are many parameters that must be considered in calculating exposure ages, and the production rates and age ranges in Table 7.1 are only estimates to be used as general guidelines for the applicability of an isotope to a particular problem.

The basic equation necessary to derive an exposure age for a radioactive cosmogenic nuclide (N), with decay constant λ, relates the amount present in a surface to the processes of production and radioactive decay:

$$(1)\quad dN/dt = \text{production} - \text{loss by radioactive decay} = P - N\lambda$$

where P is the production rate (atoms/g/yr), N is isotope concentration (atoms/g), t is time (years), and λ is the decay constant (yr^{-1}). The solution to this equation is

$$(2)\qquad N = \frac{P}{\lambda}\left(1 - e^{-\lambda t}\right).$$

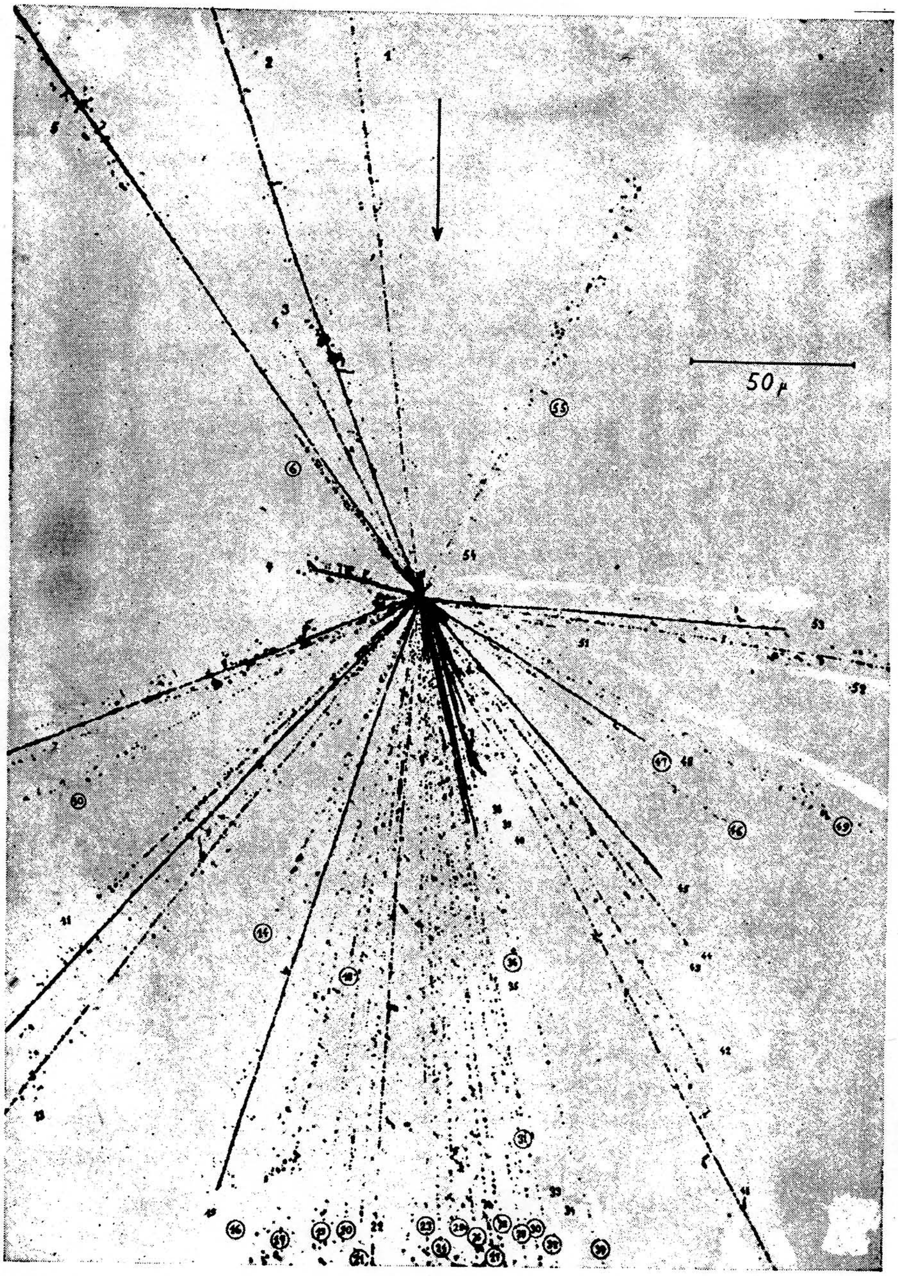

Figure 7.1. A spallation event found in an electron sensitive photographic emulsion, flown near the top of the atmosphere (15 g/cm^2) by Le Prince-Ringuet et al. (1949) (reproduced from Rossi 1952). The spallation-inducing particle is infered to be a neutron (arrow). A number of different nucleii and atomic particles are produced by this event; only the charged ones are recorded by the emulsion.

The production rate, P, depends on altitude, latitude, and the depth of the sample below the rock surface. The dependence on rock depth can be described by an exponential,

$$P = P_0 e^{-d/L} \tag{3}$$

Where P_0 is the production rate at the surface, d is the depth below the surface in g/cm^2, and L is a constant called the apparent vertical attenuation length, with units of g/cm^2. Units of g/cm^2 are universally used to normalize for density variations (conversion from these units to distance in cm requires division by the density).

Table 7.1. Some Cosmogenic Nuclides Used for Exposure Dating

Isotope	*Half-Life (years)*	*Measurement Method*	*Procedural Comments*	*Approximate Production Rate*[a] *atoms/g/yr (sea level, >55N)*	*Approximate Age Range*[b]
^{3}He	Stable	Mass Spectrometry	Diffusive loss? high production rate; lowest detection limit; inherited He	160 (olivine)	1 ka to ca. 3 Ma
^{10}Be	1.5×10^{6}	AMS	Atmospheric contamination	6 (quartz)	3 ka to 4 Ma
^{26}Al	7.16×10^{5}	AMS	^{27}Al interference (must use Al-poor minerals)	37 (quartz)	5 ka to 2 Ma
^{36}Cl	3.08×10^{5}	AMS	No mineral separates; composition-dependent (produced by spallation and slow neutrons)	8 (basalt)	5 ka to 1 Ma
^{21}Ne	Stable	Mass Spectrometry	Inherited neon; useful for old samples	45 (olivine)	7 ka to 10 Ma(?)
^{14}C	5,730	AMS	Shortest half-life; atmospheric ^{14}C contamination	20 (basalt)	1 ka to 18 ka
^{41}Ca	103×10^{3}	AMS	Useful half-life; difficult measurement	?	to 300 ka

[a] References for production rates:
^{3}He: Brook and Kurz 1993; Kurz et al. 1990.
^{10}Be: Nishiizumi, Winterer, et al. 1989.
^{21}Ne: Staudacher and Allegre 1991; Hudson et al. 1991.
^{36}Cl: Zreda et al. 1991.
^{14}C: Jull et al. 1992.

[b] Age limits calculated using detection limit of 10^{6} atoms for nuclides measured by AMS and assuming that 50 grams of sample is processed. Upper limits determined by three times the half-life.

The depth dependence is significant (L is approximately 150–160 g/cm^2, as discussed below), with production rates decreasing by a factor of two in approximately 50 cm in normal rocks; consequently, erosion may in some cases significantly affect exposure ages. A similar equation describes the dependence of the production rate on altitude:

$$P = P_0 e^{(1033-d_a)/L} \tag{4}$$

where 1033 g/cm^2 is the atmospheric depth at sea level, and d_a is the atmospheric depth in units of g/cm^2 (for example, 3 kilometers is 714 g/cm^2). The altitude can be converted to these units using standard tables, or the conversion can be made using a polynomial fit to tabulated data (for example, Lal 1991):

$$d_a = 1033 - 121.95\,(A_{km}) + 5.657\,(A_{km})^2 - 0.1095\,(A_{km})^3 \tag{5}$$

where A_{km} is the altitude in kilometers. Equation 4 is an approximation based on cosmic-ray neutron-monitor data. A slightly more complex version of altitude scaling is included in the production rate scaling factors published by Lal (1991), accounting for production by high-energy neutrons, which have a different latitude dependence than the lower-energy component. Production rates must also be scaled for the effects of the earth's magnetic field, which causes production to be somewhat higher at high latitudes than low latitudes. This can be accomplished by using relative latitude effects determined in mobile cosmic ray neutron monitor studies (see Pomerantz and Agarwal 1962; Simpson and Fagot 1956), and are also accounted for by the scaling factors given by Lal (1991). Lal's scaling, presented in polynomial form in Lal (1991), has come to be generally used by a number of workers in this field, and is discussed further below.

The effects of erosion can be included in the above equations; assuming a constant erosion rate, equation 2 becomes

$$(6) \qquad N = \frac{P}{\lambda + E/L}\left(1 - e^{-(\lambda + E/L)t}\right)$$

where E is erosion rate in g/cm^2/yr.

For a stable cosmogenic nuclide in a surface with no erosion, the above equations reduce simply to:

$$(7) \qquad N = Pt$$

and for a stable cosmogenic nuclide with constant erosion rate:

$$(8) \qquad N = \frac{PL}{E}\left(1 - e^{-(E/L)t}\right).$$

The most common issues that must be confronted by a geochronologist when attempting to date any object are the age range and accuracy of the method and its applicability to the sample in question. As discussed below, each of the cosmogenic nuclides in Table 7.1 has advantages and disadvantages, which must be considered individually. In addition surface exposure dating with any cosmogenic nuclide involves a number of common assumptions:

1. The production rates are well known;
2. The sample has been perfectly exposed to cosmic rays (no erosion or cover of any kind), or these parameters must be independently constrained;
3. The sample has not previously been exposed to cosmic rays;
4. The sample has remained a closed system during the exposure (no loss or contamination).

Each of these assumptions must be carefully considered for each geochronological situation. Accurate production rates are important for determining absolute ages, but if relative ages within a sequence of samples is the object of the study, then absolute production rates are less important. If erosion or cover has been important to a particular surface, then any measured age will be a minimum. The magnitude of under estimation is determined by the depth and time dependence of the erosion or cover. The importance of previous exposure will depend on the sample type and age range and in some cases can be evaluated based on field evidence; lava flows, for example cannot have had prior exposure to cosmic rays. If a sample was buried between exposure periods, then measurement of nuclides with different half-lives can place some constraints on the prior exposure history (see Klein et al. 1986). The "closed-system" assumption also depends on the nuclide; diffusive ^{3}He loss is an important issue for some minerals (Brook and Kurz 1993; Cerling 1990; Trull et al. 1991), and addition of atmospheric cosmogenic ^{10}Be and ^{36}Cl can be a problem for some samples (Brown et al. 1991; Zreda et al. 1991). The discussion below is intended to give background information and some examples of each of these assumptions. The reader should keep in mind that simple "dating" of surfaces or exposure ages of objects is not the only application of these techniques that may be relevant to archaeology. As the equations above show, cosmogenic nuclides are sensitive to erosion and burial processes and in favorable cases may provide useful information about erosion histories, histories of previously exposed surfaces and objects now buried and also past sediment or artifact transport in the surface environment.

Production of Cosmogenic Nuclides in Terrestrial Rocks

Interactions of Cosmic Rays with Terrestrial Material

The cosmic-ray energy at ground level is dominantly produced by galactic cosmic rays (GCR), which strike the earth from all directions and whose origin remains one of the great unsolved mysteries of physics. At the top of the atmosphere, the primary GCR flux consists dominantly of positively charged protons, with lesser amounts of alpha particles and heavy nuclei. The energies of these particles are extremely high, up to approximately 10^{19} ev. Solar cosmic rays, which originate from the sun, have significantly lower energy and are not important for *in situ* production at the earth's surface. The GCR strike atoms in the atmosphere, inducing nuclear reactions, such as the spallation event shown in Figure 7.1, which eject particles with sufficient energy to produce further nuclear reactions. By the time the resultant cosmic-ray flux reaches sea level, it is significantly attenuated, having induced many reactions in the atmosphere, and the composition of the particles has changed, so that neutrons are the dominant spallation-inducing particles.

Because the primary GCR particles are charged, the earth's magnetic field deflects them and acts as a filter, keeping out particles with energy lower than that necessary to overcome the magnetic field force vector (referred to in the literature as the "cut-off rigidity"). Because the earth's magnetic field lines are approximately parallel to the surface at the equator and perpendicular to the surface at the poles, this effect is most important at the equator and negligible at latitudes higher than approximately 55° (see Shea et al. 1987). Therefore there is a significant latitudinal variation in cosmic-ray flux and production rates. The sea level cosmic-ray flux varies by approximately 50 percent from the equator to the poles, based on mobile neutron-monitor data (Pomerantz and Agarwal 1962; Rose et al. 1956), a factor that varies with the solar cycle, because the GCR flux is modulated by the solar wind. In surface exposure dating, it is usually reasonable to assume that solar cycle variations average out over periods longer than 10 ka, and the dominant viewpoint in cosmic ray physics is that the flux to the earth has been approximately constant over the last few million years (Reedy et al. 1983).

However, due to the relationship between the earth's dipole moment, the cosmic-ray flux, and temporal changes in the earth's magnetic field, production rates of cosmogenic nuclides may vary as a function of time. It is well documented that the earth's dipole moment has varied significantly in the past (on the 10-ka time scale: McElhinny and Senanayake 1982), and hence the cosmic-ray flux at the earth's surface has also varied. This effect has long been debated with respect to calibration of the ^{14}C time scale, but it is generally recognized that variations in ^{14}C production rate are at least partly due to dipole-moment fluctuations (Damon et al. 1978). With respect to surface exposure dating, changes in production rates due to dipole moment effects are insignificant at high latitude (>55°N), but are a potential complication at low latitudes. The only experimental evidence to document the magnitude of this effect, from Hawaiian lava flows, suggests that the production rate may have varied by up to a factor of two within the last 10 ka (Kurz et al. 1990). However, Kurz et al. (1990) concluded that uncertainties in the experimental data and also in existing estimates for the relationship between dipole moment and sea-level production make it difficult to assess this quantitatively.

Production Rate Calculation

The estimation of cosmogenic nuclide production rates in the earth's atmosphere and in surficial rocks was pioneered by Lal and coworkers (Lal 1958, 1988, 1991; Lal et al. 1958; Lal and Peters 1967). They used a semiempirical approach that involved counting the spallation events in photographic emulsions flown at high altitude in the earth's atmosphere. The nuclear events recorded in such emulsions are often called "stars," based on their appearance; one such event is shown in Figure 7.1 and demonstrates that spallation generates many different nuclear fragments. Based on the number of spallation events counted in photographic emulsions, their appearance, and estimates of the branching ratio for each nuclide, Lal and Peters (1967) estimated production rates in the atmosphere for many cosmogenic nuclides. Using neutron-monitor data from around the world, they then developed curves that allow extrapolation to all latitudes and altitudes. For many isotopes these estimates can also give an approximate value for production rates in surficial rocks, and they have been used by many researchers. The star production rates derived by Lal and Peters (1967) have recently been summarized by Lal (1991) in polynomial form that can be used to scale production rates to any latitude and altitude based on the Lal and Peters (1967) formulation. Production rates can also be estimated using knowledge of nuclear reaction cross sections, target compositions, and the cosmic-ray energy spectrum and flux (Lal 1991; O'Brien 1979; Yokoyama et al. 1977). Such estimates are severely hampered, however, by the lack of appropriate reaction cross sections for neutron-induced spallation.

There are a number of limitations to the above estimates. Much of the data were collected at high altitude during the 1950s, and the latitude and altitude corrections are large. The graphical compilation of these data by Lal and Peters (1967) and the parameterization by Lal (1991) do not readily lend themselves to estimates of the uncertainties in absolute production rates or the altitude/latitude scaling (the original data are not published). Nevertheless the Lal and Peters's (1967) scaling factors for production rates are the most comprehensive and commonly used. More information can be found in Lal and Peters (1967) and Lal (1988, 1991).

Production Rate Calibration

Because production rates are critical to surface exposure dating, and because of the uncertainties in theoretical estimates, a number of workers have attempted to calibrate them (for particular nuclides) against surfaces of known age (Cerling 1990; Kurz 1986a; Kurz et al. 1990; Nishiizumi et al. 1989; Phillips et al. 1986; Zreda et al. 1991). The difficulty with this approach is in finding surfaces that have been accurately dated with other techniques and that have experienced no erosion or cover. Lava flows are ideal, because their surface morphology can be used to evaluate erosion, and there are many that have been dated by ^{14}C. The difficulty here is that the ages of the best flows are very young (<15 ka) and there are relatively small amounts of cosmogenic nuclides in these surfaces. ^{3}He, ^{36}Cl, and ^{21}Ne production rates have been calibrated this way (Cerling 1990; Kurz 1986b, 1987; Kurz et al. 1990; Poreda and Cerling 1992; Zreda et al. 1991). Other attempts to use this type of approach have included glacially polished surfaces for ^{10}Be and ^{26}Al (Nishiizumi, et al. 1989), where the age of glaciation is based on ^{14}C ages of related glacial deposits. The glacial striations and polish demonstrate that the surface has been well preserved, but any previous cover would lower the production rates, and glacial striations are typically better preserved under soil or glacial sediments.

Depth Dependence in Surface Rocks

As mentioned above, production rates vary with altitude, latitude, depth within the surface, and shielding. The variations with depth and altitude are caused by a decrease in cosmic-ray flux as it interacts with matter. The most common form of altitude and depth dependence is an exponential (equation 3 above), with reported values of the attenuation length in the atmosphere (L) between 150 and 220 g/cm^2 (units of g/cm^2 are used to eliminate variations in density). In surface materials this parameter is important, because it controls erosion rate calculations and corrections for overlying material. A number of measurements of L have been made in rock drill cores. Some of these results are summarized in Figure 7.2 (^{3}He from Kurz 1986b; ^{10}Be from Brown, Brook et al. 1992), showing that the production rates decrease by a factor of two with approximately 50 cm depth (e-folding length of approximately 60 cm), demonstrating that it is critical to document the geometry of the sample and to evaluate the impor-

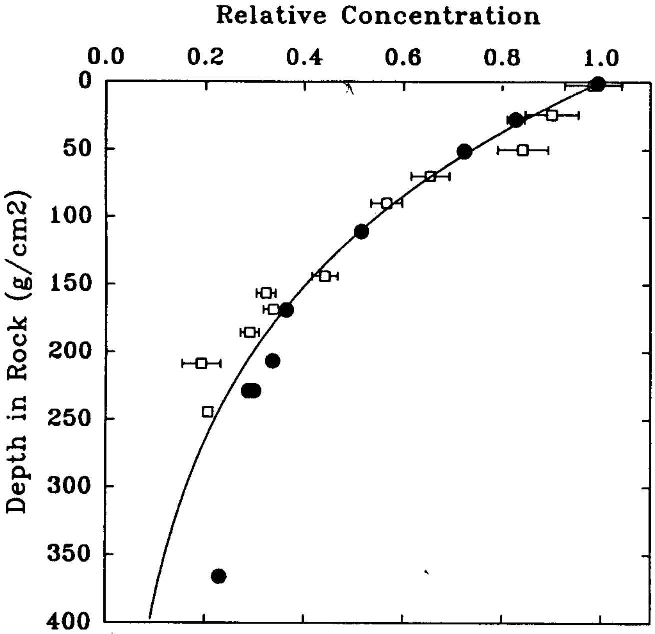

Figure 7.2. Depth profiles in surficial rocks, found by Kurz (1986b) for ^{3}He (solid circles) in a Hawaiian lava flow and Brown, Brook et al. (1992) for ^{10}Be (squares) in an Antarctic sandstone. The data were normalized to the extrapolated concentration in the respective surfaces, to allow direct comparison (that is, concentration in the top of profile defined as 1). Also shown is the profile expected from simple exponential decrease with depth (solid line). The density of both rock types is approximately 2.3 g/cm^{3}, so the total depth range shown is approximately 1.5 m.

tance of erosion and soil cover. The values obtained for L using existing drill core data are 165±7 g/cm^{2} for ^{3}He, 145 ±7 g/cm^{2} for ^{10}Be, and 156±12 g/cm^{2} for ^{26}Al (Brown, Brook, et al. 1992; Kurz 1986b); slightly higher values of L have been reported by Olinger et al. (1992) for ^{21}Ne and ^{10}Be (178 and 172 g/cm^{2}, respectively) in the Bandolier Tuff in New Mexico. Note that because these nuclides are primarily produced by neutron-induced spallation, these values will not apply to nuclides that have other production mechanisms. For example Zreda and Phillips (chapter 8 of this volume) suggest that ^{36}Cl has a more complex depth dependence than that shown in Figure 7.2, due to significant production by thermal neutrons. The production mechanisms also change significantly at greater depths in the rock, and these values are only appropriate near the surface. The deviation of the deepest ^{3}He data in Figure 7.2 from the predicted exponential may be related to an increased contribution from muon-induced reactions (that is, ^{6}Li(n,α)T → ^{3}He; Kurz 1986b). Muons are weakly interacting particles and are not as strongly depth-dependent as neutrons; consequently, below two meters depth, muon-induced reactions become more important relative to spallation (see Lal 1988).

Scaling Problems

In practice determining an absolute surface exposure age requires scaling-calibration data for the latitude and altitude of the sample in question. As an extreme example of the uncertainties that can be introduced into the calculated ages, we consider several different methods to extrapolate existing calibrations for ^{3}He production rates to high latitude. Table 7.2 (from Brook and Kurz 1993) summarizes the reported measurements of ^{3}He production rates (Kurz et al. 1990 at 20°N; Cerling 1990 at 43°N) and the different methods for scaling these data to high latitudes, in addition to available theoretical estimates. The different methods of extrapolation are based on different formulations of the relationship between altitude, latitude, solar cycle effects, and the cosmic-ray neutron flux, and can yield large discrepancies. For example the scaling of Yokoyama et al. (1977) is based on high-altitude neutron measurements (Light et al. 1973), which have a greater latitude dependence than sea level neutrons. The discrepancy between the theoretical ^{3}He production rate estimate and the experimental determinations may result from lack of knowledge of reaction cross sections for neutron-induced spallation (Lal 1991), and perhaps uncertainty regarding the sea level neutron fluxes as well. Further details regarding the table may be found in Brook and Kurz (1993). The important point illustrated in Table 7.2 is that the different formulations give significantly different production rates scaled to sea level and high altitude, and these extrapolations are an important

Table 7.2. Estimates of ^{3}He Production Rates Scaled to Sea Level and above 50° Geomagnetic Latitude, Using Four Scaling Methods Discussed in Text

	Altitude and Latitude Scaling Method[a]			
	1	*2*	*3*	*4*
Measured Production Rates				
Kurz (1987)	156±5	126±4	146±5	352±11
Kurz et al. (1990)	201±48	191±59	221±44	453±109
Cerling (1990)	153±5	153±5	116±4	216±7
Calculated Production Rates				
Lal (1991)			75[b]	
Yokoyama et al. (1977)				219[c]
Lal and Peters (1967)			150[d]	

[a] 1 = scaling based on latitude invariant vertical atmospheric attenuation length (160 g/cm^{2}) and neutron data of Pomerantz and Agarwal (1962) and Rose et al. (1956); 2 = scaling based on Lingenfelter (1963); 3 = scaling based on Lal and Peters (1967); 4 = scaling based on Yokoyama et al. (1977).

[b] Based on cosmic-ray neutron energy spectra and cross sections for proton interactions, calculated for quartz.

[c] Based on ^{3}H production rate, calculated using cross sections for proton interactions and ^{3}H/^{3}He production ratio of Lal and Peters (1967).

[d] Based on cosmic-ray star production/isotope production relationship of Lal and Peters (1967), with a likely uncertainty of approximately 20 percent (Lal, personal communication).

source of uncertainty for numerical-age calculations. In addition the altitudinal and latitudinal variability of production rates may also be energy-dependent, such that cosmic-ray neutron monitor data need corrections to be relevant to nuclide production (Lal 1958; Lal and Peters 1967).

One approach to the problem of altitude and latitude scaling of production rates is to measure production rates in dated surfaces at different altitudes (the latter approach assuming a constant cosmic-ray flux). Where this has been done at one latitude (see Brown et al. 1991; Zreda et al. 1991), results are consistent with the altitude scaling of Lal and Peters (1967), but additional work will be necessary to confirm this conclusion. Another promising approach is to measure cosmogenic nuclides, in experimental targets set out over long periods of time (see Mabuchi et al. 1971). Ultimately this approach, along with results from surfaces of known age, will remove the uncertainties related to altitude and latitude scaling. At present the scaling based on Lingenfelter (1963) or Lal and Peters (1967) seem to be the most comprehensive data available, but they yield significantly different results. Adequate understanding of the scaling problem and recognition of the inherent uncertainties are necessary for the interpretation of *in situ* cosmogenic nuclide data.

^{3}He and ^{21}Ne

^{3}He and ^{21}Ne are both stable nuclides, which is an advantage for dating older surfaces. In addition they can both be measured with a conventional magnetic sector mass spectrometer. However, they also share the disadvantage that a correction must be applied for ^{3}He and ^{21}Ne present in the rock prior to exposure to cosmic rays (referred to here as the "inherited component"). At present there is considerably more data for ^{3}He regarding this problem, and the correction can be made in a relatively simple manner in many cases.

Although the presence of cosmogenic helium in terrestrial rocks was only discovered in 1985, there is now a considerable amount of data for various rock types (Anthony and Poths 1992; Brook and Kurz 1993; Brook et al. 1993; Cerling 1990; Craig and Poreda 1986; Kurz 1986a, 1986b; Kurz et al. 1990; Porcelli et al. 1987). Virtually all of the existing cosmogenic ^{3}He data for igneous rocks is from olivine and clinopyroxene phenocrysts, due to significant loss of ^{3}He from the finer-grained basaltic groundmass (Kurz 1986a). The helium contained within basaltic phenocrysts is a mixture of cosmogenic and magmatic helium. The quantity of cosmogenic ^{3}He can be calculated as follows:

(9) $^3\text{He (cosmogenic)} = {}^3\text{He (total)} - {}^3\text{He (inherited)}.$

The quantity of inherited ^{3}He can be calculated as follows:

(10) $^3\text{He (inherited)} = {}^4\text{He (total)} \times ({}^3\text{He}/{}^4\text{He})$ (inherited).

This calculation is valid because only a negligible amount of cosmogenic ^{4}He is produced; that is, the ^{3}He/^{4}He ratios in igneous rocks are typically ca. 10^{-5} and the ^{3}He/^{4}He ratio produced by spallation is ca. 0.2 (Mazor et al. 1970). The inherited ^{3}He/^{4}He ratio is determined by crushing *in vacuo*, which selectively releases the inherited component (dominantly contained by fluid and melt inclusions). The cosmogenic ^{3}He is not released by crushing olivine and clinopyroxene (Kurz 1986a, 1986b), but is released by melting the previously crushed sample.

It is important to note that the simple correction for inherited helium given by equations 9 and 10 does not necessarily apply to other rock types. Brook and Kurz (1993) performed experiments on quartz from Antarctic sandstones and found that significant amounts of cosmogenic ^{3}He were released by crushing in a vaccum, thus demonstrating that the formulation above does not apply to quartz. In this instance they assumed that all the inherited helium is radiogenic, with ^{3}He/^{4}He ratios of ca. 10^{-8} (Andrews 1985; Morrison and Pine 1955), which in this case resulted in a small correction for rocks older than 100 ka. However, it should be emphasized that more data are necessary to evaluate these assumptions for other rock types, and the assumptions regarding inherited helium must be carefully considered.

Due to the low detection limits for ^{3}He (5,000 to 10,000 atoms) and the fact that ^{3}He has the highest production rates of any cosmogenic nuclide (see Table 7.1), helium measurements can be applied to extremely young surfaces, as young as 1,000 B.P. at sea level and potentially younger at higher elevations. Mineral separations are difficult for some rock types and can limit the applicability of the technique, depending on the amount of sample that must be processed, which in turn depends on crystal abundances, detection limits, age, and altitude. For samples older than 1,000 B.P. at sea level, roughly 200 mg of olivine or clinopyroxene is typically adequate for ^{3}He analysis.

The first attempts to calibrate production rates and to test errors in the method were made on a number of Hawaiian lava flows (Kurz 1986a, 1986b; Kurz et al. 1990). These lava flows were dated by ^{14}C (see Rubin et al. 1987), using charcoal from beneath the flows, and they are ideal for ^{3}He dating because olivine is abundant. The published Hawaiian data are summarized in Figure 7.3; they are noteworthy in that they demonstrate that the ^{3}He method can be used for very young ages, in this instance the youngest samples are 1,000 B.P. in age. There is a reasonable correlation between ^{14}C and ^{3}He; both methods yield the same stratigraphy. However, there are some notable deviations from the concordance line in Figure 7.3. This could be partly due to an incorrect production rate (see discussion above), because the concordance line is based on a ^{3}He production rate of 125 atoms/g/yr, derived from the youngest samples because the surfaces are best preserved (note that a best fit to the data yields a lower production rate). Kurz et al. (1990) suggested that some of the deviation, particularly for those samples between 2,000 and 7,000 B.P., could be related to modulation of the cosmic-ray flux by the earth's magnetic field. The decrease in production rate (that is, lower ^{3}He ages) during this period would be consistent with archaeomagnetic evidence for a strong dipole moment (McElhinny and Senanayake 1982) and recent suggestions of strong nondipole fields in Hawaii at this time (Manki-

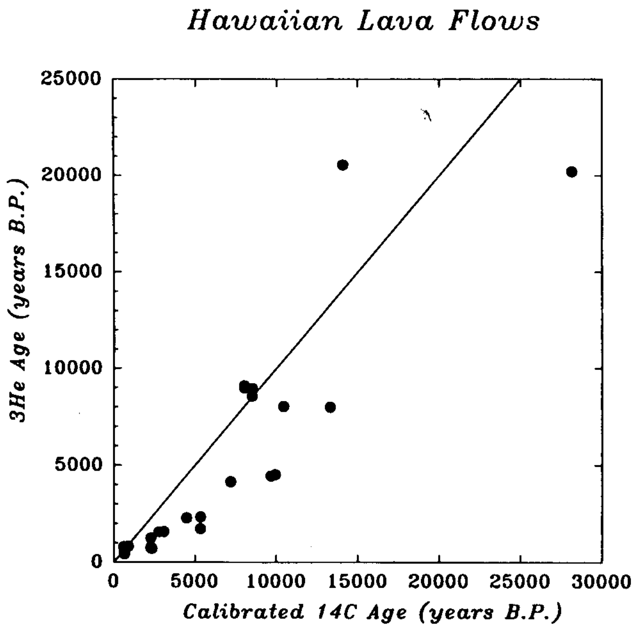

Figure 7.3. Calculated ^{3}He exposure ages from Hawaiian radiocarbon-dated lava flows (assuming a sea level production rate of 125 atoms/g/yr) compared to the ^{14}C ages for the same flows. The helium data are from Kurz et al. (1990), and the radiocarbon dates from Rubin et al. (1987). Note that the line is not a best fit to the data, but is based on the ^{3}He production rate from the youngest samples. The use of a lower ^{3}He production rate would yield better agreement for the samples below the line, but no single production rate explains all the data. One possible explanation is that production rates have fluctuated in the past, due to changes in the earth's magnetic field (Kurz et al. 1990).

nen and Champion 1992). Soil cover and erosion would also lower apparent ^{3}He ages, but Kurz et al. (1990) suggested that these processes are more likely for the older samples. This suite of samples illustrates some of the uncertainties in the method and suggests that the uncertainty in absolute ages is typically 30 percent for this age range. Cerling's (1990) production rate (see Table 7.2) from a single lava flow erupted into a dated shoreline in glacial lake Bonneville is within the error of Kurz et al. (1990) rates, but additional tests will be required to refine production rates and reduce the uncertainties.

Another important issue for using ^{3}He in exposure age dating is diffusive loss from mineral grains. Helium diffusion is extremely slow in olivine and clinopyroxene, making this effect unimportant on time scales less than 10 Ma at surficial temperatures (Hart 1984; Trull et al. 1991). Cerling (1990), based on several attempts to measure helium in four quartz samples, suggested that loss rates are too rapid from this mineral to allow its use for exposure dating. In contrast, Trull et al. (1991) performed diffusion measurements on Antarctic quartz samples that suggest extremely slow diffusive loss. Extrapolation of their diffusion data, which were collected at temperatures in excess of 100°C, to surficial temperatures, suggested that helium loss would be insignificant for time scales less than 2 Ma (Trull et al. 1991). In addition Brook et al. (1993) have reported extremely old ^{3}He exposure ages from Antarctic glacial moraines, discussed further below, which demonstrate that significant quantities of cosmic-ray-produced helium are retained in quartz. However, several aspects of the helium data indicate that the older samples have experienced some loss. Brook and Kurz (1993) demonstrated that the concentration of cosmogenic ^{3}He in quartz grains from a single sample increases significantly with grain size, suggesting a diffusive loss mechanism. Both ^{3}He/^{10}Be and ^{3}He/^{21}Ne ratios in these same samples suggest approximately 40 to 50 percent helium loss in the 2–3 Ma samples (Brook and Kurz 1993; Brown et al. 1991; Staudacher and Allegre 1991). Although the evidence suggests helium losses from quartz over million-year exposure periods, there is also evidence that diffusive loss from quartz is insignificant for shorter exposure periods. Based on the ages obtained from Arena Valley, and particularly the correlation between ^{3}He and ^{10}Be for the younger samples (see Figures 7.4 and 7.5), it seems that ^{3}He will be a useful technique for quartz samples younger than 400 ka. The discrepancy between this conclusion and that of Cerling (1990), who suggested that helium is lost from quartz on time scales shorter than 10 ka, may relate to an inappropriate correction for inherited helium (that is, equation 10) by

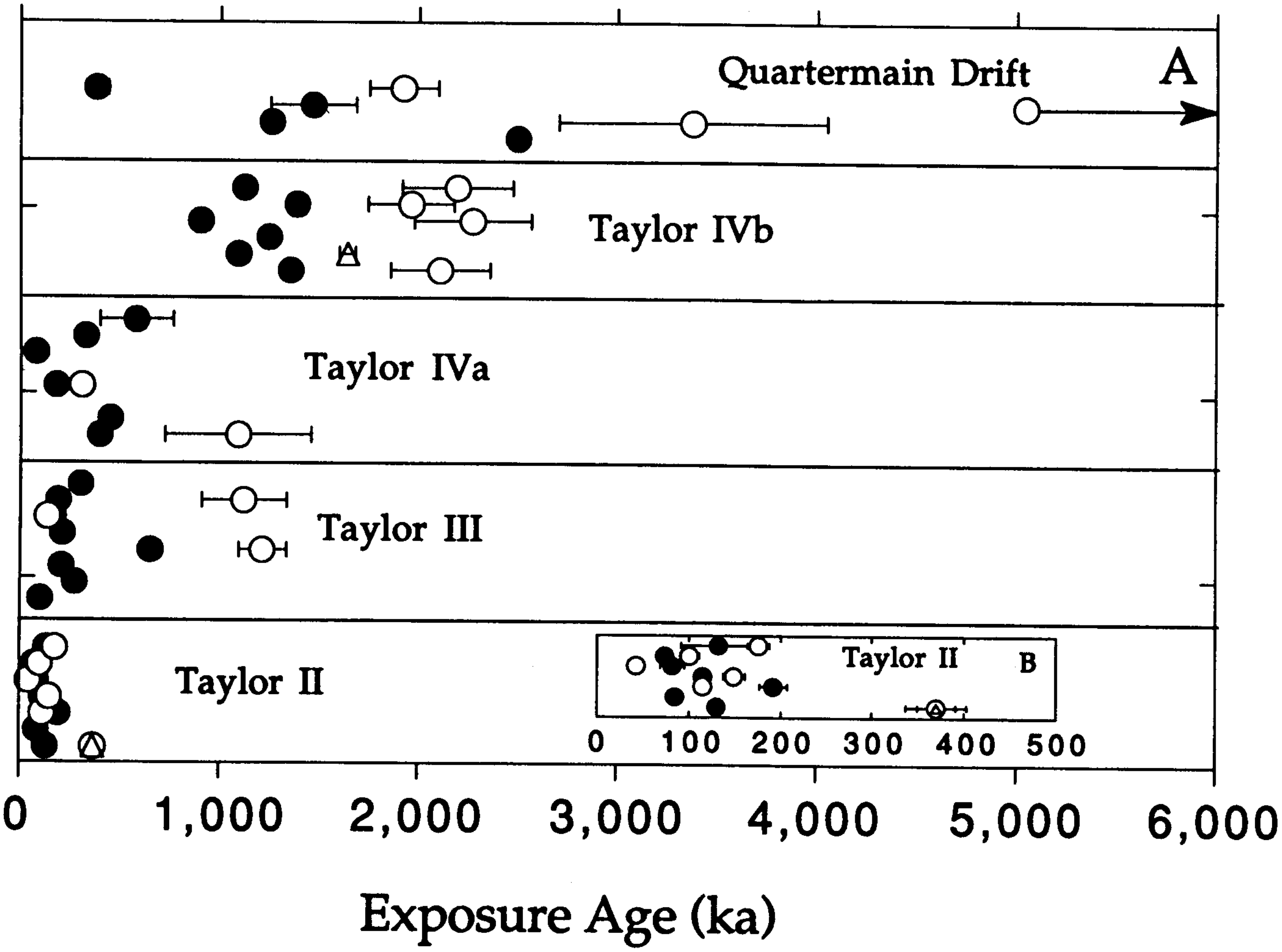

Figure 7.4. ^{3}He and ^{10}Be age distributions for a set of glacial moraine boulders in Arena Valley, Southern Victoria Land, Antarctica (Brook, Brook, et al. 1993). The solid circles are ^{3}He ages, and the open circles are ^{10}Be ages; most of the samples are quartz grains from quartz sandstone boulders. The data demonstrate that a single moraine can display a range of exposure ages, requiring detailed sampling of such features. The discrepancy between ^{3}He and ^{10}Be ages for the older samples is probably related to helium loss from samples older than 500 ka. The inset gives an expanded view of the ages from the youngest moraine; the agreement between ^{3}He and ^{10}Be ages suggests little loss of ^{3}He in 100 ka.

Cerling (1990), or to differences in sample mineralogy, grain size, or environmental conditions.

^{21}Ne has significantly slower diffusion rates from minerals and may be extremely useful, particularly in samples that have undergone helium loss (Graf et al. 1991; Marti and Craig 1987; Staudacher and Allegre 1991). As with ^{3}He, a correction must be made for inherited neon. In young volcanic rocks this correction for magmatic ^{21}Ne is analogous to corrections for magmatic ^{3}He. In older rocks a correction also must be made for ^{21}Ne produced by (α,n) reactions with thermal neutrons (for example, ^{24}Mg(n,α)^{21}Ne). The importance of the correction depends on the exposure age, the uranium and thorium concentraton (these elements

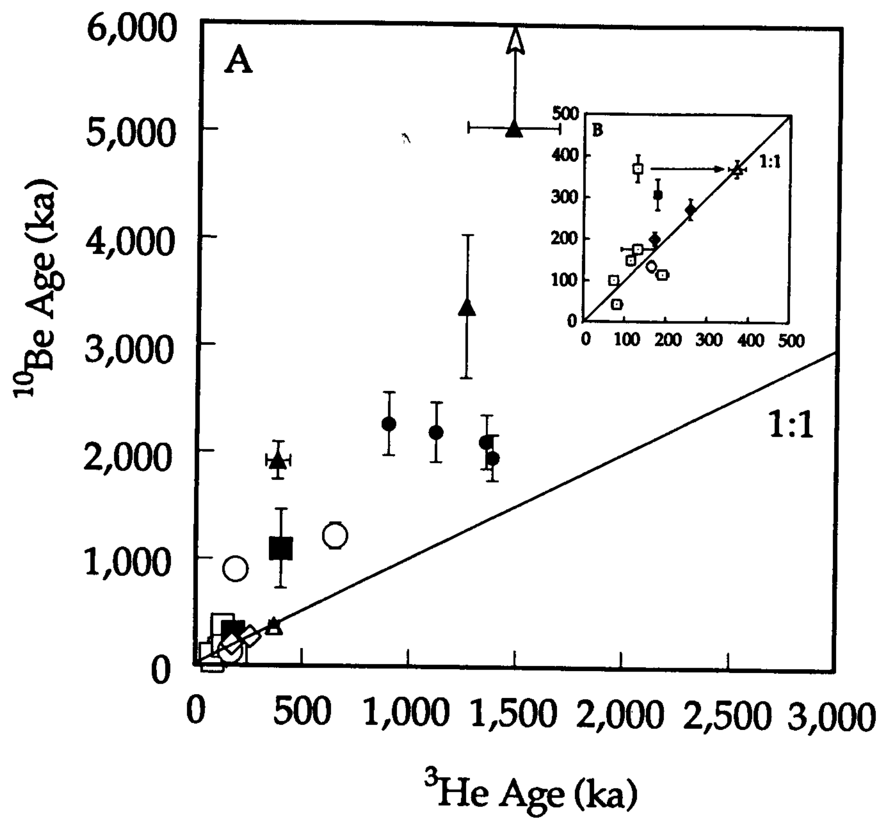

Figure 7.5. ^{3}He and ^{10}Be ages for a subsample of the samples shown in Figure 7.4, calculated with ^{3}He and ^{10}Be production rates of 191 and atoms/g/yr, respectively (Brook et al. 1993). Different symbols indicate the different moraines (see Figure 7.4). Also as demonstrated by Figure 7.4, ^{3}He ages are younger than ^{10}Be ages, due to ^{3}He loss for the older samples. However, the younger samples suggest good agreement between ^{3}He and ^{10}Be ages for samples younger than 400 ka (see inset). The arrow in the inset connects two different grain size fractions from the same sample; the larger grains fall close to the concordance line.

produce alpha particles), and the concentrations of targets for (α,n) reactions. Because neon has three isotopes, discriminating different neon components may be simpler than for helium.

Recently Poreda and Cerling (1992) have undertaken the calibration of ^{21}Ne production rates in olivine and plagioclase in young (< 20 ka) volcanic rocks, finding values of about 45 and 17 atoms/g/yr, respectively, at sea level and high latitude. The difference is due to the different compositions of the minerals. The composition dependence of spallation production has been well studied in meteorites (Eugster 1988; Vogt et al. 1990), and Poreda and Cerling (1992) also studied the compositional dependence of ^{21}Ne production in olivine. More work of this type is needed, but the demonstration that ^{21}Ne can be used in a number of mineral phases is important. For samples with old crystallization ages and short exposure ages, the correction for inherited neon (that is, preexposure, nucleogenic ^{21}Ne) will be important.

Antarctica has been the focus of a number of exposure dating studies (Brown et al.1991; Brook et al. 1993; Cerling 1990; Nishiizumi et al. 1986; Nishiizumi, Kohl, Arnold, et al. 1991; Staudacher and Allegre 1991). In a number of ways, Antarctica is an ideal environment for this technique, because it is a cold desert and erosion is very slow (Brown et al. 1991; Nishiizumi, Kohl, Arnold, et al. 1991). In addition the glacial history of the continent is critical to understanding climatic history, and there are few other methods for dating the glacial deposits. Figure 7.4 shows ^{3}He exposure ages from glacial moraines in Arena Valley, which is adjacent to Taylor Glacier in the Dry Valleys region of Southern Victoria Land (Brook et al. 1993). The moraines record the advance and retreat of Taylor Glacier, which is an outlet glacier from the East Antarctic Ice

Sheet, and have been the subject of a number of glaciological studies (such as Denton et al. 1989). The data shown in Figure 7.4 are from quartz sandstone moraine boulders and are in good agreement with prior attempts at relative ages of the moraines. Even though the older quartz samples have experienced some ^{3}He loss, as discussed above, they have ^{3}He ages as old as 1.5 Ma. A comparison of ^{3}He with ^{10}Be ages, shown in Figure 7.5, suggests that the loss has been approximately 50 percent of the ^{3}He for the oldest samples (Brook et al. 1993; Brown et al. 1991). Despite the evidence of loss, the stratigraphy obtained using the two techniques is generally consistent.

The data shown in Figures 7.4 and 7.5 illustrate the importance of sampling to any successful application of surface exposure dating. In the case of the moraine boulders, the assumptions of surface exposure dating require that the boulders have not rolled or been covered in the past, and that they have not experienced prior exposure to cosmic rays. The data shown in Figure 7.4 demonstrate that single samples are not adequate to characterize a single moraine. In addition it is important to use several different techniques, rather than limit age determination to one cosmogenic nuclide, because each of the methods provides unique information.

^{10}Be and ^{26}Al

^{10}Be and ^{26}Al are often measured together, because they complement each other in many ways. Both require AMS for measurement because of their low abundances and the presence of isobaric interferences (for example, ^{10}B for ^{10}Be and ^{26}Mg for ^{26}Al). Almost all of the ^{10}Be and ^{26}Al data for terrestrial surface exposure dating are for the mineral quartz, because it has a simple target chemistry, is a ubiquitous mineral in nature, is chemically resistant, and has very low stable aluminum concentrations (typically ca. 100–1000 ppm). The latter two attributes are important from a laboratory standpoint. Contamination of grain surfaces with ^{10}Be produced in the atmosphere can be significant (Brown et al. 1991); the use of quartz allows rigorous acid leaching to remove the surface contamination, while leaving interior material intact. In fact this approach has proven extremely successful in eliminating this potential contaminant (Brown et al. 1991; Nishiizumi et al. 1989). In addition, because the quartz matrix is not susceptible to chemical weathering, alteration of the mineral during its exposure history does not affect the results. Because it is not produced in significant quantities in the atmosphere, ^{26}Al does not suffer from similar contamination problems, but the amount of ^{27}Al (the stable isotope) in the sample must be low to obtain reasonable counting statistics for ^{26}Al. For both ^{10}Be and ^{26}Al, the analysis is carried out as a simple isotope dilution. ^{9}Be and ^{27}Al are added to the samples as carriers, and the 26/27 and 10/9 ratios are measured using the AMS. For ^{10}Be analysis of quartz, there is generally not enough naturally occurring ^{9}Be to be of concern, but for ^{27}Al this is not the case (because aluminum is a major constituent of most minerals), and the stable aluminum concentration must be determined prior to chemical purification; then additional ^{27}Al can be added, if necessary. The amount of natural stable aluminum that can be tolerated depends on sample size and the amount of ^{26}Al present, but concentrations greater than a few thousand ppm will make analysis difficult, if not impossible, for most samples. Therefore aluminosilicate minerals, such as feldspars and micas, are unsuitable for ^{26}Al analysis. However, quartz can be separated from other minerals in some rocks by selective acid leaching (see Kohl and Nishiizumi 1992).

The half-lives of ^{10}Be and ^{26}Al (1.5 Ma and 720 ka) are suitable for using the isotopes as a pair, as they both can be used to determine ages of Quaternary deposits. By measuring both isotopes, the internal consistency of the data can be evaluated, since the results should be concordant, and erosion rate constraints are considerably more reliable with two isotopes.

There has been only one calibration of the ^{26}Al and ^{10}Be production rates, by Nishiizumi, Winterer, et al. (1989), who studied glacially polished surfaces on the Sierra Nevada Mountains of California. Their results indicated production rates (at ca. 44° geomagnetic latitude) of 62±3 and 374±28 atoms/g/yr for ^{10}Be and ^{26}Al, respectively (at 3.34 km, or 685 g/cm^2, altitude). This corresponds to production rates of 6 and 37 atoms/g/yr at sea level. These production rates are based, however, on a somewhat uncertain glacial chronology, and the authors indicate additional uncertainties of approximately 10 percent. Another important outcome of this work was the determination of the ^{26}Al/^{10}Be production ratio (ca. 6), a useful parameter that is not greatly affected by altitude or latitude. With improvement in analytical precision, dating based on isotope pairs may be more useful than those based on the absolute concentration of a single isotope, because the use of ratios reduces uncertainties associated with absolute production rates, erosion, and skyline shielding (Brown, Brook, et al. 1992).

^{10}Be and ^{26}Al analyses are possible in other mineral phases, particularly olivine, where some work has been done (Nishiizumi, Klein, et al. 1990), although obtaining enough mineral separate to make an adequate measurement (at least several grams are necessary) is difficult. Calibration of ^{10}Be and ^{26}Al production rates should be possible with lava flow samples, but any comparison of such data with the results for quartz requires understanding the effect of target chemistry on production rates. Although this is fairly well understood in meteorites, it is not clear that it is adequately understood yet in terrestrial rocks. Further work in this area will be necessary. Because ^{10}Be and ^{26}Al are radioactive (with half-lives of 1.5×10^6 and 725,000 yrs respectively), their dating range is limited by their half-lives (see Table 7.1). The lower limits of the potential age range are a function of latitude, altitude, burial depth, and sample size. The practical upper limit is defined by approach to "saturation," where production balances decay and erosional loss.

One aspect of the cosmogenic radionuclides, not yet discussed in this article, is the possibility of using concentrations to place constraints on erosion rates. Over time isotope concentrations reach a steady state with respect to erosion and radioactive decay. The length of time necessary to reach this steady state depends on the half-life of the nuclide and the erosion rate, and if steady state is assumed, a maximum erosion rate can be calculated by setting t equal to infinity in equation 6, yielding:

$$E = (P/N - \lambda)L. \tag{11}$$

If it is clear that a steady state has been reached, this yields an average erosion rate; otherwise the rate is a maximum value. The ^{26}Al–^{10}Be pair can be useful in this regard; if both yield the same erosion rate, then one can be confident that the samples are at or near steady state. From equation 11, however, it should be obvious that the erosion rate calculation depends on accurate knowledge of production rates, which will be an important source of uncertainty, particularly for low erosion rates. It should also be recognized that these calculations give erosion rates only at particular points on individual surfaces, and using these data to characterize erosion on regional or continental scales may be misleading, depending on the nature of the surfaces sampled, as shown recently by Brown, Stallard et al. (1992). In addition equation 11 assumes a constant erosion rate. More complex scenarios, including variable erosion rates and spalling of discrete thickness of the surface, can be modeled (Lal 1991).

Most of the geological applications of ^{10}Be and ^{26}Al to terrestrial geochronology have been demonstrations of the potential of the technique for determining exposure ages and erosion rates of glacial landforms in Antarctica (for example, Brook et al. 1993; Brown et al. 1991; Nishiizumi, Kohl, Shoemaker, et al. 1991). Earlier work on Libyan desert glass (Klein et al. 1986) demonstrated the power of the technique, and recent work on Meteor Crater in Arizona (Nishiizumi, Kohl, Shoemaker, et al.1991; Phillips et al. 1991) demonstrated excellent agreement between ^{10}Be, ^{26}Al and ^{36}Cl methods. Recent examples of work in Antarctica will demonstrate some of the potential and problems involved in using ^{26}Al and ^{10}Be for surface exposure chronology. Nishiizumi, Kohl, Arnold, et al. (1991) reported ^{10}Be and ^{26}Al data for a number of quartz samples from the Transantarctic Mountains and Alan Hills regions of Antarctica. The data indicated minimum exposure ages ranging from 36 ka to greater than 4 Ma, and they were able to use the data to calculate maximum erosion rates of roughly 10^{-5} cm/yr for their oldest samples. A major result of this work has been the quantification of very low erosion rates of exposed rocks in Antarctica. These low erosion rates make surface exposure dating particularly useful for polar desert regions, since old surfaces of geochronologic interest are so well preserved.

We have initiated efforts to use ^{10}Be and ^{26}Al in conjunction with ^{3}He data, to determine the ages of surficial glacial deposits in the Dry Valleys region of Antarctica, and to use these data to study the Plio-Pliestocene history of the East and West Antarctic ice sheets (Brook, et al. 1993; Brown et al. 1991). Initial efforts have concentrated on dating boulders in Quaternary moraines in the Dry Valleys region, deposits that have been well studied by other workers, who have been hampered by lack of detailed chronology (for example, Bockheim 1992; Denton et al. 1989). One moraine system in Arena Valley, a small hanging valley intruded into in the past by the Taylor Glacier, an outlet glacier of the East Antarctic Ice Sheet, has been studied in detail (Brook et al. 1993; Brown et al. 1991). There are difficulties in dating individual boulders, because they all may have somewhat different exposure histories; but when enough data are accumulated, coherent sets of ages appear to emerge (Brook, et al. 1993; Brown et al. 1991; Phillips et al. 1990; Zreda and Phillips, chapter 8 of this volume) that are consistent with independent evidence. The statistical treatment of such data remains an important question (see also Zreda and Phillips, chapter 8 of this volume). Brook, et al. (1993) chose to take the mean of the measurements shown in Figure 7.4 as the best estimate of the true exposure age of the deposits, while Zreda et al. (1991) assume that maximum ages are the most reliable ones, at least for older deposits. These studies have also verified the low erosion rates reported earlier for many Antarctic samples (Brown et al. 1991; Nishiizumi, Kohl, Arnold et al. 1991). Brown et al. (1991) also developed a statistical treatment of ^{10}Be and ^{26}Al data that indicated that the production rates determined by Nishiizumi, Winterer, et al. (1989), based on a 10-ka exposure age, were consistent with the average rate over the last 10^6 years (with fairly large uncertainties).

The multiple-isotope approach has been useful in these studies. For example potential problems for ^{3}He loss and atmospheric contamination of ^{10}Be can be evaluated by measuring several isotopes in the same samples (for example, Brook et al. 1993; Brown et al. 1991), greatly facilitating the interpretation of age distributions (Figure 7.4), as well as providing new information about the isotope systematics of these nuclides.

From the above discussion, the most important factor in choosing *in situ* ^{10}Be or ^{26}Al for a surface exposure problem relates to the mineral phases present in the material to be dated. Existing studies are primarily confined to quartz, due to the relative ease in obtaining pure mineral separates in large quantities ($>$ 1g), but may be extended to other low-aluminum mineral phases. Glass might also be studied in an archaeological context, although large sample sizes would probably be required. Apart from dating surfaces, other applications of these nuclides are chiefly in estimating the erosion rates of surfaces and sedimentary processes. The latter application has been described in principle by Lal (1987) and employed by Klein et al. (1986) and Nishiizumi, Kohl, Arnold, (1991). The burial and exposure history of individual sedimentary particles or units, as well as the origin of soil or weathering horizons, can be constrained. For example "disequilibrium" in the ^{10}Be–^{26}Al isotopic system can indicate periods of exposure followed by burial, because ^{26}Al and ^{10}Be decay at different rates, producing a burial signal (Klein et al. 1986). The exponential attentuation of production with depth in rocks can be also be employed in studying sedimentary sections. Exposure of material prior to deposition could be distinguished from exposure after deposition by the presence or absence of the exponential decrease. This type of work may prove useful in understanding the timing of deposition versus exposure or determining the origin of sedimentary horizons (deposition versus weathering).

^{14}C and ^{41}Ca

Preliminary work with these two isotopes suggests some interesting and useful applications. Their chief attraction is their relatively short half-lives (5,730 yr for ^{14}C and 103 ka for ^{41}Ca; Klein et al. 1991), resulting in greatest utility for processes on time scales of 0–40 ka and 0–500 ka, and a greater sensitivity to erosion and burial on these time scales than is the case for ^{3}He, ^{21}Ne, ^{10}Be, ^{26}Al, or ^{36}Cl. As are the other spallation-produced nuclides, ^{14}C is produced by the spallation of major elements in a silicate matrix. The production rate should be composition-dependent, and although it has been detected in several terrestrial samples (Jull et al. 1992), ^{14}C production rates have not been systematically calibrated. These measurements are difficult, due to atmospheric contamination and the small quantities of ^{14}C present, but Jull et al. (1992) were able to suggest that previous age estimates (from longer-lived cosmogenic nuclides) for some of their samples were incorrect, because the ^{14}C data suggest significant erosion not resolved by longer-lived cosmogenic nuclides. Clearly more work must be done, but *in situ* ^{14}C dating should prove extremely useful in the future.

^{41}Ca is produced in terrestrial rocks by thermal neutron capture by ^{40}Ca, but *in situ* ^{41}Ca measurements in terrestrial rocks have not, to our knowledge, been reported in the literature. A number of ^{41}Ca measurements have been made with the intention of dating bones, although consistent results have not yet been obtained (Fink et al. 1990; Raisbeck and Yiou 1979). There are a number of difficulties with ^{41}Ca measurements, most notably that the ^{41}Ca/^{40}Ca ratios in most samples are quite low, less than 10^{-14} (Fink et al. 1990). Present AMS backgrounds are at best 5×10^{-16}, so data quality is probably limited at present. Since the half-life of ^{41}Ca is so attractive for studying Quaternary events, progress in ^{41}Ca applications and measurement would be extremely useful.

Summary

It is clear from the discussion presented above that more calibrations of the production rates and of the scaling for altitude and latitude are necessary to reduce the uncertainties in the technique. Calibrations of this kind are underway in a number of laboratories, and it is likely that the precision of surface exposure will therefore improve significantly in the near future.

As with any dating technique, surface exposure dating requires attention to detail in sample collection, laboratory procedure, and evaluation of assumptions. Because the application of the technique is still in its infancy, much of the discussion given before focuses on the outstanding problems. In any application of the technique, careful sample collection and geological constraints on erosion and soil cover are critical. Nevertheless, surface exposure dating is already well enough developed to be of great utility. A good example of the success of the technique is the concordance between several different age determinations on the surfaces of Meteor Crater (Nishiizumi, Kohl, Shoemaker, et al. 1991; Phillips et al. 1991). With the wide range of cosmogenic nuclides given in Table 7.1, surface exposure dating will find many new applications in fields such as archaeology, geomorphology, glacial geology, and volcanology.

Acknowledgments

The authors gratefully acknowledge the collaboration of R. Ackert, G. Denton, D. Lott, D. Kammer, W. Jenkins, E. Brown, G. Raisbeck, and F. Yiou. Some of the work discussed here, and the writing of this manuscript, were supported by grants EAR91–06820 and DPP91–17458 from the National Science Foundation. Erik Brown and Greg Ravizza provided helpful editorial and scientific comments on the manuscript. We also thank Charlotte Beck for her encouragement and patience in the face of adversity.

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