Using in situ produced cosmogenic isotopes to estimate rates of landscape evolution: A review from the geomorphic perspective

Paul R. Bierman
Department of Geology, University of Vermont, Burlington

Abstract. The application of in-situ produced cosmogenic isotopes to problems in geomorphology has increased rapidly over the past decade. At least 57 papers and numerous abstracts have been published since the mid-1980s when the first mass-spectrometric measurements of terrestrially produced cosmogenic isotopes were made. Taken at face value, these studies provide quantitative information about rates of landscape evolution and landform age; however, the significance of calculated erosion rates and exposure ages depends strongly on the models used to interpret isotopic data, the validity of assumptions inherent to these models, and the geologic surroundings in which the samples were collected. This paper attempts to place cosmogenic isotope studies in a geomorphic context by reviewing fundamentals of the method and evaluating the validity of assumptions under which these data have been interpreted. At present, the establishment of high-precision, cosmogenically based glacial and alluvial chronologies is stymied by the evolution of geomorphic surfaces, the erosion of rock from sampled boulders, the potential for isotope inheritance from previous exposure, and the uncertainty of isotopic measurements. Uncertainties in isotope production rates and the observed variability of exposure ages on individual geomorphic surfaces limit the confidence with which cosmogenic ages can be correlated reliably with those obtained by other techniques. Estimation of erosion rates at single points on the landscape gives useful small-scale information. Extrapolation of these rates over longer time and larger spatial scales is less sure and most likely biased toward lower erosion rates by the inadvertent selection of resistant sample sites. However, because erosion rates are so poorly constrained at present, even estimates to within a factor of 2 may be of significant value to geomorphologists and tectonicists.

Introduction

Deciphering the shape and history of Earth's surface, in the context of tectonic and climatic forces, is a fundamental objective of geomorphology. In order to understand landscape evolution and the denudation of continents, one must not only identify germane physical processes, one must also know the time at which features of the landscape formed and/or the rate at which these features change.

Recent advances in analytical chemistry and nuclear physics have provided geomorphologists with the opportunity to constrain rates of landscape evolution directly. Use of high-sensitivity noble gas and accelerator mass spectrometers (AMS) now allows quantitative abundance measurement of extremely rare isotopes including those produced by the interaction of cosmic rays with rock and soil [*Raisbeck et al.*, 1983; *Elmore and Phillips*, 1987]. For example, the abundance in near-surface rocks of terrestrially produced cosmogenic isotopes, such as ^{3}He, ^{10}Be, ^{21}Ne, ^{26}Al and ^{36}Cl, has been used along with interpretive models to evaluate rates of landscape change near the proposed Yucca Mountain waste repository [*Whitney and Harrington*, 1993], to correlate alpine moraines with continental glaciations recorded in the deep-sea ^{18}O record [*Phillips et al.*, 1990], to test models of desert pavement formation [*Wells et al.*, 1991] and to suggest that outcrops in Antarctica have lost only decimeters from their surfaces during millions of years of subaerial exposure [*Nishiizumi et al.*, 1991a)].

In a general sense, the abundance of in situ produced cosmogenic isotopes is proportional to landscape stability and/or age. Favorable circumstances allow isotope abundances to be used as a proxy for erosion rate or exposure age, both of which express the residence time of the sampled material near Earth's surface. It is important to recognize that all such estimates are model ages or model erosion rates; they depend strongly on the model used to interpret isotope abundance and the validity of the model assumptions.

This paper reviews the theory and application of in situ produced cosmogenic isotopes specifically from a geomorphologic point of view. It investigates the validity of assumptions inherent in interpretive models and considers isotope studies in the context of what is known about the rate and distribution of germane surface processes. In conclusion, the paper presents current limitations of the method and speculates on the usefulness of cosmogenic isotope measurements for evaluating the rate at which landscapes change and topography evolves.

Terrestrial Production of Cosmogenic Isotopes

A variety of isotopes, termed cosmogenic, are produced as cosmic rays interact with rock and soil. Six of these isotopes (^{3}He, ^{10}Be, ^{14}C, ^{21}Ne, ^{26}Al, and ^{36}Cl) are now measured routinely and have been used in geomorphic studies since the late 1980s. Isotope production rates are very low, ranging from <10 to >100 atoms g^{-1} yr^{-1} at sea level and high latitude (Table 1).

Paper number 94JB00459.
0148-0227/94/94JB-00459$05.00

1435

Table 1. Commonly Used In Situ Produced Cosmogenic Isotopes

Isotope	Half-Life λ	Production Rate,* atoms g^{-1} yr^{-1} (Sea Level, >60°)	Production Rate Studies	Primary Cosmogenic Reaction(s) [*Lal*, 1988]	Radiogenic Formation	Measurement Technique (Compound)	Common Sample Material
^{3}He	stable	(silicate) 191 (olv) 70-220 (olv) 140 (olv) 143 (bas) 94 (bas) 73 by composition (olv) 109	*Brook and Kurz* [1993] *Kurz et al.* [1990] *Cerling* [1990] *Kurz* [1986b] *Craig and Poreda* [1986] *Lal* [1987] *Lal* [1991] *Poreda and Cerling* [1992]	spallation of O, Mg, Si, Fe	$^{6}Li\,(n,\alpha)\,^{3}H$	MS	olivine, pyroxene, quartz
^{10}Be	1.5 m.y. (4.6×10^{-7} yr^{-1})	(qtz) 6.5 (qtz) 10 (qtz) 6.0 (qtz) 6.4 (olv) 5.25	*Lal and Arnold* [1985] *Middleton and Klein* [1987] *Nishiizumi et al.* [1989] *Brown et al.* [1991] *Nishiizumi* [1991]	spallation of O, Mg, Si, Fe	$^{7}Li\,(\alpha,p)\,^{10}Be$	AMS (BeO)	quartz, olivine
^{14}C	5730 years (1.2×10^{-6} yr^{-1})	(qtz) 17.5 (silicate) 19	*Lal* [1991] *Jull et al.* [1992]	spallation of O	$^{11}B\,(\alpha,p)\,^{14}C$ $^{17}O\,(n,\alpha)\,^{14}C$	AMS (C)	whole rock, quartz
^{21}Ne	stable	(qtz) 16.6 by composition (qtz) 0.4 x ^{26}Al (olv) 0.29 x ^{3}He (olv) 0.4 x ^{3}He (olv) 45, (plag) 16	*Hudson et al.* [1991] *Lal* [1991] *Graf et al.* [1991] *Staudacher and Allegre* [1993b] *Poreda and Cerling* [1992] *Poreda and Cerling* [1992]	spallation of Mg, Al, Si, Fe	$^{18}O\,(\alpha,n)\,^{21}Ne$	MS	olivine, quartz, plagioclase
^{26}Al	0.7 m.y. (9.9×10^{-7} yr^{-1})	(qtz) 27.5 (qtz) 70 (qtz) 36.8 (qtz) 41.7 (olv) 15.4	*Lal and Arnold* [1985] *Middleton and Klein* [1987] *Nishiizumi et al.* [1989] *Brown et al.* [1991] *Nishiizumi* [1991]	spallation of Si, Al, Fe	$^{23}Na\,(\alpha,n)\,^{26}Al$	AMS (Al_2O_3)	quartz, olivine
^{36}Cl	0.3 m.y. (2.3×10^{-6} yr^{-1})	2600-3000 atom mol Ca^{-1} yr^{-1} 3500-8500 atom mol K^{-1} yr^{-1} 1.2 x 10^{6} atom mol Cl^{-1} yr^{-1}	*Davis and Schaeffer* [1955]; *Yokoyama et al.* [1977]; *Zreda et al.* [1991]; *Swanson et al.* [1993]; *Phillips et al.* [1993]	spallation of K, Ca, Fe $^{35}Cl\,(n,\gamma)\,^{36}Cl$	$^{35}Cl\,(n,\gamma)\,^{36}Cl$	AMS (AgCl)	whole rock

*Corrected to sea level and high latitude by *Lal* [1991] if data were uncorrected by original author; qtz, quartz; olv, olivine; bas, basalt; silicate, silicate rock; plag, plagioclase; by composition indicates that the author has provided a formula for calculating production rates based on mineral composition.

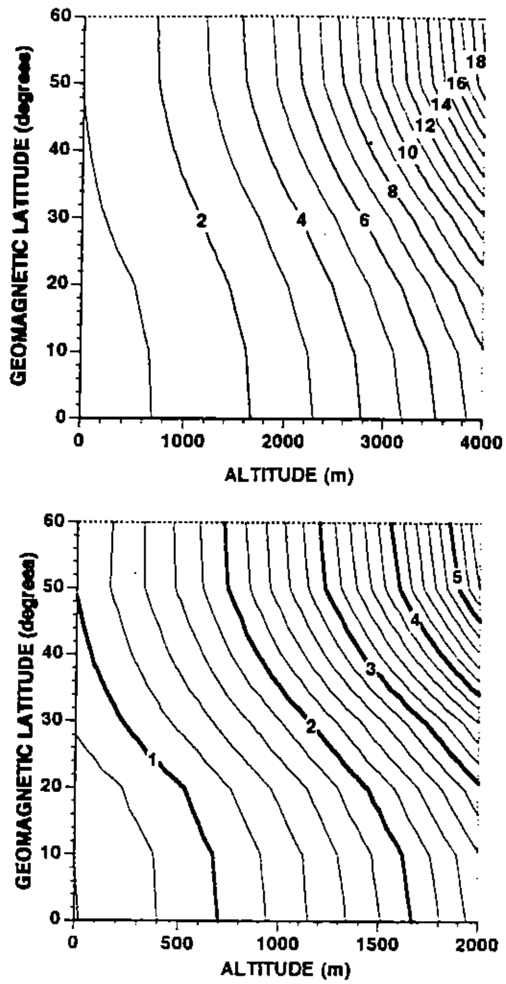

Figure 1. Contour plots of altitude/latitude correction factors for cosmic ray flux and isotope production at Earth's surface based on data presented by *Lal* [1991]. (top) Altitude range 0-4000 m; (bottom) range of 0-2000 m. To obtain production rate at any altitude or latitude, multiply sea-level, high-latitude production rate in Table 1 by factor read from the chart.

Flux of Cosmic Rays

Cosmic rays, predominately neutrons, continually bombard Earth's surface and produce cosmogenic isotopes. These neutrons are secondary particles resulting from the interaction of primary cosmic rays (mostly protons) with Earth's atmosphere. At Earth's surface, the flux of cosmic rays, and consequently the rate of isotope production, depends largely on the modulation of the primary galactic cosmic-ray flux by Earth's magnetic field and modulation of the secondary flux by Earth's atmosphere. Primary cosmic rays (protons) are deflected more strongly by the magnetic field at the equator than at the poles. The secondary neutrons interact with atmospheric gasses so that the resulting flux is attenuated with increasing atmospheric depth. As a result, cosmogenic isotope production will be greatest in samples exposed at high elevation and high latitude.

On the basis of contemporary neutron measurements, *Lal* [1991] describes the altitude/latitude variation of the cosmic ray flux and resulting isotope production with a third-order polynomial (Figure 1). He estimates that these corrections are accurate to within +/- 10%. Most workers use Lal's method and data for scaling isotope production rates, although others have been published and used [*Rose et al.*, 1956; *Pomerantz and Agarwal*, 1962; *Lingenfelter*, 1963; *Yokoyama et al.*, 1977].

Mechanisms and Rates of Cosmogenic Isotope Production

Isotopes termed "cosmogenic" are produced primarily by four types of interactions between particles and target nuclei: spallation, muon capture, neutron activation, and alpha particle interaction. Different production pathways dominate at different depths below Earth's surface. *Fabryka-Martin* [1988] provides a useful and well presented review of nuclide production mechanisms. During spallation, the target nucleus is split by the incoming particle, usually a high-energy, fast cosmic ray neutron. For example, fast neutron spallation produces ^{26}Al from silicon and ^{10}Be from oxygen. Most spallation reactions occur within the upper meter or two of Earth's surface. Muons, charged particles of low mass, can interact with the nucleus of the target atom. For instance, interactions between muons and ^{40}Ca and ^{28}Si produce ^{36}Cl and ^{26}Al, respectively. Muons interact less strongly with matter than fast neutrons; therefore muon-induced isotope production rates are lower and muon penetration depths greater than those characteristic of neutron spallation. Other isotopes, such as ^{36}Cl and ^{41}Ca, are produced primarily or in part by neutron activation, the nuclear absorption of low-energy or thermal neutrons, e.g., ^{36}Cl from ^{35}Cl and ^{41}Ca from ^{40}Ca. These thermal neutrons result either from the interaction and slowing of fast neutrons and muons or, indirectly, from the decay of U and Th [*Feige et al.*, 1968; *Andrews et al.*, 1989]. In rock and soil, alpha particles are produced primarily by the decay of U, Th, and their daughter nuclides. Alpha particles are responsible directly and indirectly for the radiogenic (or noncosmogenic) production of "cosmogenic" nuclides. For example, the absorption of an alpha particle and subsequent emission of a neutron by an ^{18}O nucleus produces ^{21}Ne in what is termed an α, n reaction. Such reactions of alpha particles with light elements are the primary source of radiogenic thermal neutrons which can produce noncosmogenic ^{36}Cl and ^{41}Ca [*Feige et al.*, 1968; *Andrews et al.*, 1989]. Most of the cosmogenic nuclides produced in situ in rock and soil are also produced by nuclear reactions in the atmosphere. These atmospherically produced nuclides are referred to as meteoric or garden variety [*Nishiizumi et al.*, 1986], and in most cases, atmospheric production rates exceed those in rock by orders of magnitude. A variety of pretreatments have been used to remove atmospherically produced isotopes from samples in order to isolate those produced in situ [*Leavy*, 1987; *Nishiizumi et al.*, 1989; *Brown et al.*, 1991; *Kohl and Nishiizumi*, 1992].

Isotope Production Below Earth's Surface

Understanding the depth distribution of cosmogenic isotope production is important for deciphering cosmogenic isotope abundances both in terms of erosion rates and exposure ages. Erosion rate models require quantification of isotope production rates experienced by a parcel of rock as it moves from depth toward Earth's surface during erosion. Many rocks sampled for exposure dating have probably lost mass from their surface [*Blackwelder*, 1927; *Birkeland*, 1984; *Bierman and Gillespie*, 1991]. The effect of such mass loss on measured isotope abundance and model ages depends on the relationship between the rate of isotope production and depth.

In general, rates of in situ cosmogenic isotope production are highest at Earth's surface and decrease rapidly with depth as incoming cosmic ray particles are slowed and absorbed. In rock (ρ = 2.7 g cm^{3}), roughly half the fast cosmic ray neutrons are absorbed above a depth of 45 cm. For isotopes produced primarily by fast neutron spallation (^{3}He, ^{10}Be, ^{14}C, ^{21}Ne, ^{26}Al), the production rate (P_x) at depth (x) in a material of given density (ρ) can be described well, over the first several meters of rock, with an exponential expression considering the surface production rate (P_O):

$$P_x = P_0 \, e^{-(x\rho/\Lambda)} \qquad (1)$$

The characteristic attenuation length (Λ) for fast neutrons is between 150 and 170 g cm^{-2} and has been shown to be roughly invariant at different locations on Earth's surface [*Kurz*, 1986b; *Lal*, 1991; *Brown et al.*, 1992; *Sarda et al.*, 1993].

For isotopes such as ^{36}Cl and ^{41}Ca, which are produced in significant quantities by thermal neutron activation, the situation is more complex because the distribution of thermal neutrons with depth is not well modeled by an exponential expression [*Bierman*, 1993]. Because fast neutrons scatter isotropically as they interact with matter, some backscatter away from the air/rock interface before they are slowed to thermal energies [*O'Brien et al.*, 1978]. Consequently, isotope production by thermal neutrons actually increases between the surface and a depth of about 15 cm resulting in the possible overestimation of ^{36}Cl exposure ages if samples are inadvertently collected from eroding surfaces.

Isotope Production Rates

Rates of cosmogenic isotope production vary with time, location, and target composition. For most isotopes, production rate estimates remain somewhat uncertain and are constrained by only a few field experiments and theoretical calculations. Theoretical calculations rely on probabilities of nuclear interactions and measurements of the cosmic ray flux. Empirical estimates are field-based calibrations during which isotope abundances are measured in samples for which exposure ages have been estimated by other means. Field calibration of isotope production rates over the 100,000-year time scale is confounded by the paucity of well-dated geomorphic surfaces that have neither eroded nor been buried since their initial exposure.

On the basis of abundance measurements of short-lived isotopes and probabilities of nuclear interactions, [*Yokoyama et al.* [1977], *Lal* [1988], and *Lal* [1991] calculate production rates for a variety of long-lived isotopes. Field studies have recently been used to calibrate the production of ^{3}He [*Cerling*, 1990; *Kurz et al.*, 1990], ^{21}Ne [*Hudson et al.*, 1991], ^{10}Be and ^{26}Al [*Nishiizumi et al.*, 1989], and ^{36}Cl [*Zreda et al.*, 1991]. Recent work supports earlier estimates of ^{10}Be and ^{26}Al production rates [*Brown et al.*, 1991; *Nishiizumi et al.*, 1991a] but there are conflicting estimates of ^{36}Cl production rates [*Yokoyama et al.*, 1977; *Swanson et al.*, 1992; *Phillips et al.*, 1993; *Swanson et al.*, 1993]. It is possible that some production rate discrepancies, such as those for ^{3}He [*Cerling*, 1990; *Kurz et al.*, 1990], may be due to or reconciled by the use of different protocols for altitude/latitude correction [*Brook and Kurz*, 1993].

There is both direct and indirect evidence that magnetic field intensity and therefore isotope production rates have changed significantly and irregularly over the Pleistocene Epoch [*McFadden and McElhinny*, 1982; *Bard et al.*, 1990; *Kurz et al.*, 1990; Lal, 1991; *McElhinny and Senanayake*, 1982; *Tric et al.*, 1992]. *Reedy et al.* [1983] suggest changes in atmospheric ^{10}Be production of +/-30% over 100 kyr and +/-10% over periods of 200 kyr. There is also evidence for short periods of exceptionally high rates of isotope production [*Raisbeck et al.*, 1985].

Because rocks integrate cosmic ray exposure, the effect of changing production rates will be damped over time. Calculations based on paleointensity data of *Tric et al.* [1992] suggest that if an accurate, average production rate could be determined for the past 30 kyr, production rate variations would result in age deviations no larger than 16% for Holocene samples and less than 10% for Pleistocene samples [*Sarda et al.*, 1993]. However, some production rate calibrations are based on samples younger than 15 kyr [*Nishiizumi et al.*, 1989; *Cerling*, 1990; *Kurz et al.*, 1990; *Zreda et al.*, 1991], a period of higher than average field strength and thus lower than average isotope production. In addition, data of *Kurz et al.* [1990] suggest nearly a twofold change in time-integrated ^{3}He production rates over the past 13 kyr, a greater change than one would predict by considering records of magnetic field paleointensity [*Kurz et al.*, 1990; *Mazaud et al.*, 1991]. Currently, the magnitude of the uncertainty introduced by changing production rates, and by calibration on relatively young surfaces, is not well constrained but appears to be of the order of 10-20% for Holocene and latest Pleistocene samples.

Sample Collection, Preparation, and Isotopic Analysis

Sample collection and preparation can be time consuming. Samples are frequently collected using hammers and chisels to exploit preexisting weaknesses in the substrate. Mechanical coring devices can also be used, although they require that cooling water be available on site. Interpretation of isotope abundance is simplest if samples are collected from large, flat-lying surfaces. Samples collected from surfaces having a more complex geometry such as vertical faces or spherical boulders require the calculation of effective neutron fluxes [*Nishiizumi et al.*, 1989; *Staudacher and Allegre*, 1993a]. Such calculations are more difficult and uncertain for thermal neutrons [*Zreda et al.*, 1993].

The amount and complexity of sample preparation, prior to analysis, differ for each cosmogenic isotope. Noble gas measurements (He and Ne) are made on mineral grains using mass spectrometry. Extensive chemical pretreatment and preparation of a chemically pure target are required for isotopic analysis of C, Be, Al, and Cl by AMS. Initial sample size is usually less than 500 g of rock but depends on a variety of factors, including lithology, machine sensitivity, and exposure time.

Noble gas analyses usually require that one mineral or phase be separated either by hand-picking, magnetic properties, or density. The size of the aliquot usually ranges from 0.1 to 1 g. For ^{3}He analyses, the minerals pyroxene and olivine have been used most commonly and with the greatest success [*Craig and Poreda*, 1986; *Kurz*, 1986a; b; *Kurz et al.*, 1990; *Cerling*, 1990; *Wells et al.*, 1991; *Anthony and Poths*, 1992]. Several research groups have presented data which suggest that quartz does not retain ^{3}He quantitatively in most terrestrial environments [*Cerling*, 1990; *Graf et al.*, 1991; *Hudson et al.*, 1991].

Cosmogenic ^{10}Be and ^{26}Al have been measured in quartz and olivine [*Lal and Arnold*, 1985; *Nishiizumi et al.*, 1986; *Klein et al.*, 1986; *Nishiizumi et al.*, 1989; *Nishiizumi*, 1991; *Nishiizumi et al.*, 1991a; b; *Brown et al.*, 1991, 1992]. In order to measure cosmogenic ^{26}Al reliably, the mineral phase must contain low levels of the stable isotope, ^{27}Al; quartz is ideal as it usually contains less than 100 ppm of Al. For Be measurements, the most important characteristic is the ability to prepare a sample free of atmospherically produced ^{10}Be. Again, quartz has been shown to be useful because it is resistant to weathering and alteration and can be effectively cleaned of meteoric ^{10}Be by HF etching [*Nishiizumi et al.*, 1989; *Brown et al.*, 1991; *Kohl and Nishiizumi*, 1992]. Chemical preparation of Be and Al targets for AMS involves dissolution of the mineral phase by HF.

Cosmogenic ^{36}Cl and ^{14}C have been measured in a variety of lithologies without mineral separation. Interpreting ^{36}Cl abundances measured in whole rock samples requires knowing the major and trace element composition of the rock. Chlorine has been extracted from rock samples by a variety of means including fusion, dissolution by HF and precipitation, air stripping from solution [*Leavy*, 1987; *Zreda et al.*, 1991] and water leaching [*Bierman*, 1993]. Whole rock analyses are not necessary for interpreting in situ produced ^{14}C abundances because, to the first approximation, ^{14}C production rates are independent of composition [*Jull et al.*, 1992]. In situ produced cosmogenic ^{14}C can be removed from rock or mineral separates by acid dissolution (D. Lal, personal communication, 1993) or fusion under oxygen [*Jull et al.*, 1992].

Accelerator mass spectrometry for ^{26}Al, ^{10}Be, and ^{36}Cl is typically able to measure isotopic ratios as low as several parts in 10^{-15}. The precision (1σ) of AMS measurement for these isotopes, based on counting statistics and reproducibility of standards, is typically < 5-7%. Nobel gas measurements typically have absolute uncertainties of $\geq$ 5% (1σ) although ratios can be measured more precisely [*Staudacher and Allegre*, 1993a].

Interpreting Isotope Abundance

Translating measured isotope abundances into geomorphically useful data requires conceptual and mathematical models built on a variety of potentially uncertain and, in some cases, untestable assumptions. Cosmogenic erosion rates and exposure ages are often quoted with uncertainties based only on the counting statistics of the underlying isotopic measurements. The true uncertainties of model ages and model erosion rates are greater, as they must encompass not only uncertainties in production parameters resulting from Earth's changing magnetic field but also model assumptions such as no-inheritance or steady state erosion.

Other Sources of "Cosmogenic" Isotopes

Isotopes termed "cosmogenic" can also be produced by a variety of means not directly related to the present period of cosmic ray exposure. If these other sources of isotope production are not considered and corrected for, isotope inventories and exposure ages will be overestimates and calculated erosion rates will be underestimates. As mentioned previously, atmospheric production of ^{10}Be, ^{14}C, and ^{36}Cl is significant; interpretive models assume that sample pretreatment has removed any meteoric contamination.

Radiogenic isotope production. The production of "cosmogenic" isotopes resulting from the decay of U and Th can, in some cases, be significant. In most cosmogenic isotope studies, no correction has been made for radiogenic production, a reasonable approach for samples exposed at high altitude or for long periods of time.

On the basis of whole rock elemental abundances, *Sharma and Middleton* [1989] conclude that radiogenic production of ^{10}Be is unimportant but that radiogenic production of ^{26}Al is significant. However, because U is concentrated in minerals other than quartz [*Tieh and Ledger*, 1981; *Turner*, 1993], radiogenic ^{26}Al is probably insignificant in many samples analyzed for cosmogenic isotopes. *Brook and Kurz* [1993] suggest that radiogenic production of ^{3}He is minimal in most rocks and that radiogenic ^{3}He can be corrected for during analysis. In situ production of ^{14}C appears to be inconsequential [*Zito et al.*, 1980; *Fabryka-Martin*, 1988; *Raisbeck and Yiou*, 1990].

Estimates of radiogenic ^{36}Cl production have been made by a number of authors [*Andrews et al.*, 1986; *Bentley et al.*, 1986; *Leavy et al.*, 1987; *Fabryka-Martin*, 1988; *Andrews et al.*, 1989]. For young samples, exposed at low elevations, radiogenic production can account for significant portion of the measured ^{36}Cl. For example, *Bierman* [1993] estimated ratios of ^{36}Cl/Cl resulting from U- and Th-induced neutrons for 129 granitic rocks using the method of *Andrews et al.* [1989] and found a mean radiogenic ^{36}Cl/Cl ratio of 21+/-18 x 10^{-15} with a range of 4 x10^{-15} to 125 x 10^{-15}.

Neon 21 can be produced radiogenically as ^{18}O absorbs alpha particles. The abundance of radiogenic ^{21}Ne can be estimated from the concentration of U and Th [*Ozima and Podosek*, 1983]. Calculations, and measurements made on shielded samples, show significant amounts of radiogenic ^{21}Ne should be produced in quartz extracted from older granites. For instance, we found a ^{21}Ne excess (over atmosphere) of 1.5 x 10^{7} atoms g^{-1} in shielded (675 g cm^{-2}; 2.5 m rock) samples from a Precambrian granite (>1 b.y.) in Texas (B. Hudson, Lawrence Livermore National Laboratory, unpublished data, 1992).

Geologic inheritance. Isotope abundance has been measured in rocks collected from a variety of landforms including moraines and alluvial or debris flow fans. When calculating model ages for such deposits, it has been assumed that boulders were deposited on the landform with no preexisting inventory of cosmogenic isotopes. There are only sparse data by which to test the assumption of no inheritance. A boulder riding on the ice of Taylor glacier, Antarctica, has a ^{3}He exposure age of 9 ka [*Brook and Kurz*, 1993]; the implications of this age and the problem of inheritance are discussed by *Brook et al.* [1993]. *Bierman* [1993] sampled several boulders in a roadcut through a moraine in the Bishop Creek drainage, east side the Sierra Nevada, California. These boulders, shielded by >8 m of till, have ratios of ^{36}Cl/Cl between 50 and 100 x 10^{-15}, of which only about half can be explained by radiogenic production.

Some isotopes, such as ^{36}Cl and ^{26}Al, have important muon production pathways and are produced in significant abundances several meters below the ground surface. As a result, one would suspect that the likelihood of boulders inheriting ^{36}Cl or ^{26}Al might be higher that the likelihood of inheriting ^{14}C from a prior exposure. For instance, at a depth of 3 m in rock, the rate of isotope production from muon stopping is still almost 50% of its maximum value [*Charalambus*, 1971; *Fabryka-Martin*, 1988; *Bilokon et al.*, 1989]. *Bierman* (1993) measured ^{36}Cl in bedrock samples shielded by several meters of rock and found ^{36}Cl/Cl ratios of 50 - 270 x 10^{-15}, several times the ratio supported radiogenically. This finding implies the potential for significant ^{36}Cl inheritance if boulders had previous exposure histories within several meters of Earth's surface. Stable (^{3}He and ^{21}Ne) or long-lived (^{10}Be) nuclides should be more susceptible to inheritance than short-lived nuclides.

Model Exposure Ages and Erosion Rates

The measured abundance of a cosmogenic nuclide will reflect production integrated through time reduced by decay in the case of an unstable nuclide. If the sampled surface were exposed rapidly and has not eroded since exposure, isotope abundance will be a function of exposure age and isotope production rate. If the sampled surface has been eroding continually since initial exposure, and initial exposure was several half-lives ago or several meters of rock have eroded since exposure, then the isotope concentration will be controlled by the erosion rate. These are two end-members in a continuum of plausible sample histories. In reality, many geologic samples have probably suffered episodic erosion after initial exposure.

On the basis of field observations it is usually difficult to determine which of the end-member models is most appropriate for evaluating isotopic data [*Bierman and Gillespie*, 1992]. Some features, for example, glacial striae, are good indications that a rock surface has not lost mass since initial exposure, although it may be difficult to assess whether unconsolidated material such as till, soil, or ash once covered the striated surface. Although steady erosion is more likely for such substrates as saprolite, the assumption of steady erosion is less valid for those bedrock surfaces which lose mass in discrete slabs >10 cm in thickness [*Lal*, 1991].

Exposure ages. The exposure age for a surface sample is calculated from the measured isotope abundance (N) after a correction has been made for radiogenic and geologic backgrounds (B) and the decay (λ) of the nuclide if it is unstable [*Lal*, 1988]:

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

Radiogenic backgrounds can be calculated from the abundance of U, Th, and the abundance of suitable target atoms. Geologic backgrounds are more uncertain and more likely to vary from sample to sample. It is assumed that the isotope production rate (P) is representative of the time frame (t) over which the exposure occurred and that no erosion has occurred since initial exposure.

Erosion rates. The abundance of a cosmogenic isotope in a surface sample can also be interpreted as a steady state erosion rate (ε). Such an approach has been considered in some detail [*Lal and Peters*, 1967; *Lal*, 1986; 1988; 1991; *Nishiizumi et al.*, 1991a; 1993]. For the case of the spallation-produced isotope, the following model is applicable. If isotope production as a function of depth is not well described by an exponential (^{36}Cl), the analytic solution (equation 3) will not accurately estimate erosion rates:

$$N = \frac{P}{\varepsilon\rho\Lambda^{-1} + \lambda} + B. \qquad (3)$$

Implicit in the derivation and application of this equation is the assumption of steady or high-frequency, periodic erosion and a constant rate of isotope production.

Exposure and erosion. In many cases it is possible to measure the concentration of more than one cosmogenic isotope in the same sample. If multiple isotopes have been measured, it should, in principle, be possible to solve both for erosion rate and for initial exposure age. Assuming constant production rate and steady erosion, *Nishiizumi et al.* [1991a] present the following model for surface samples:

$$N = \frac{P}{(\rho\varepsilon\Lambda^{-1} + \lambda)}(1 - e^{(\rho\varepsilon\Lambda^{-1}+\lambda)t}) + B \qquad (4)$$

Considering the current uncertainties in production rates and the measurement of isotope abundance, solutions of this equation are relatively imprecise and of limited use in many geologic situations [*Gillespie and Bierman*, 1991; *Nishiizumi et al.*, 1991a].

There are, however, other ways in which dual isotope measurements can be used to constrain qualitatively exposure histories [*Lal*, 1991]. For example, measurements of ^{26}Al and ^{10}Be in the same samples have been used to constrain the exposure and burial history of bedrock surfaces in the Dry Valleys of Antarctica [*Nishiizumi et al.*, 1991a]. Because the half-life of ^{26}Al is less than that of ^{10}Be, the ratio of $^{26}Al/^{10}Be$ can be interpreted as a function of erosion rate, time of burial after cosmic ray exposure or exposure age [see *Nishiizumi et al.*, 1991a, 1993].

Geomorphic Applications

Geomorphic problems which have been addressed using cosmogenic isotopes fall into three categories: estimating exposure ages of landforms, determining the erosion rate of rock surfaces, and the testing of geomorphic process models.

Model Exposure Ages

Accurate and precise dating of surface features could be useful for constraining the evolution of young volcanic fields, the timing of climate change, and rates of tectonic displacement. So far, cosmogenic isotope samples have been collected and analyzed for the purpose of constraining the age of alpine and continental glaciation [*Phillips et al.*, 1990; *Brown et al.*, 1991; *Brook et al.*, 1993], the age of volcanic features [*Cerling*, 1990; *Staudacher and Allegre*, 1993a] and the timing of cratering events [*Nishiizumi et al.*, 1991b; *Phillips et al.*, 1991].

Ironically, it is the evolution and erosion of landforms which most significantly hampers cosmogenic dating. For example, *Cerling* [1990] measured ^{3}He abundance in several Pliocene lava flows in southeastern California. He calculated exposure ages only 4-12% of the K/Ar cooling ages (2.1 and 3.7 Ma) and so concluded that the flows have probably eroded at a model rate of several meters per million years. In this case, cosmogenic isotope abundance was controlled by erosion rate rather than exposure age. Conversely, in order to calculate production rates of ^{26}Al and ^{10}Be, *Nishiizumi et al.* [1989] sampled only striated bedrock surfaces thus assuring that little rock had eroded after initial exposure. Such an approach is only viable for the length of time that rock surfaces remain exposed at the surface without significant degradation as 1-3 mm of erosion is capable of removing striae. In our experience, few, if any, continuously exposed, striated surfaces are older than the latest Pleistocene ~ 20 ka [*Bierman and Gillespie*, 1991; 1992]. Moreover, our observations and those of others (D. Dethier, personal communication, 1993) indicate that striated surfaces are best preserved below a mantle of till. These observations suggest that some, and perhaps many, striated surfaces have not been continuously exposed since formation. The effect on exposure ages of such shielding will depend on the thickness and persistence of the now-eroded till cover.

Other situations are less clear cut. In the Sierra Nevada of southeastern California, *Phillips et al.* [1990] used the abundance of ^{36}Cl and the assumption of no erosion to calculate exposure ages for 30 samples collected from boulders on several moraines at Bloody Canyon. Their published model ages are likely too high because production rates for ^{36}Cl have been revised upward [*Phillips et al.*, 1993; *Swanson et al.*, 1993]. More importantly, the cosmogenic ages they calculate for older (pre-latest Wisconsinan) moraines are at odds with a distinct stratigraphic (crosscutting) relationship [*Bursik and Gillespie*, 1993]. Specifically, samples collected from boulders on the stratigraphically oldest (Mono Basin) moraine do not have the oldest exposure ages. While it is possible that the crosscutting relationship is not as clear-cut as it appears, it is more likely that boulders on the stratigraphically older Mono Basin moraine have lower than stratigraphically expected exposure ages because they were exposed by erosion long after the moraine was deposited. The moraine appears to be quite degraded morphologically with a wide and rounded crest. Few boulders crop out on this moraine and those that do are heavily weathered.

The disagreement at Bloody Canyon between ^{36}Cl and relative, stratigraphic ages may result both from the lowering of moraine surfaces and the degradation of morainal boulders. Lowering of geomorphic surfaces will result in sequential exposure of initially buried boulders and should generate a wide distribution of exposure ages. If the rate of surface lowering is episodic or age-dependent, it is possible that boulders now exposed on older moraines could have lower exposure ages than those exposed on stratigraphically younger moraines. The destiny of exposed boulders is less certain, but the common observation that boulder frequency decreases with the age of a geomorphic surface implies that individual boulders are not well preserved over time. Loss of mass and consequently isotopes from boulder surfaces will in most cases, with the possible exception of ^{36}Cl, result in cosmogenic model ages which underestimate the age of the landform [*Bierman and Gillespie*, 1991; *Bierman*, 1993]. Unfortunately, only if isotope inheritance is be shown to be unimportant can the maximum boulder age be confidently interpreted as a minimum limiting age for the landform. *Brook et al.* [1993] provide an insightful discussion of the uncertainties inherent in the interpretation of exposure age data from moraines.

The moraine data of *Phillips et al.* [1990], *Brown et al.* [1991] and *Brook et al.* [1993] are sobering to those attempting to determine precisely, using in situ produced cosmogenic isotopes, the age of a geomorphic surface. These studies show a wide variability (5-50%, 1σ) in measured exposure ages on what have been interpreted as single moraine surfaces and some of these studies show that the variability of isotope abundance, in percentage terms, increases with surface age. Unfortunately, such exposure age variability is not unique to glacial moraines. *Bierman* [1993] finds significant boulder variability (50-100%, 1σ) using ^{36}Cl to measure exposure ages of debris flow fan surfaces in Owens Valley. Several authors find variability (43%, ^{26}Al, n=11; 41%, ^{10}Be, n=11; 13%, ^{36}Cl, n=5; all 1σ) between samples collected from ejecta blocks at Meteor Crater in Arizona [*Nishiizumi et al.*, 1991b; *Phillips et al.*, 1991; *Jull et al.*, 1992]. At Meteor Crater, maximum model exposure ages for ^{26}Al, ^{10}Be and ^{36}Cl all closely match each other and an independently determined thermoluminesence (TL) age for the cratering event (49+/- 3 ka). However, one of the five ^{36}Cl model ages and six of the 11 ^{10}Be and ^{26}Al model ages are significantly less than the TL age, implying that ejecta boulders have idiosyncratic geologic histories which may include erosion and/or delayed exposure after bolide impact. In this case, it is reasonable to accept the maximum exposure age as the minimum limiting age of the cratering event because the chance for prior cosmic ray exposure of sampled boulders was minimal.

Such distinctions are less certain and more arbitrary when exposure ages of individual samples are used to estimate depositional ages for landforms such as moraines, terraces, and alluvial fans. The difficulty in interpretation is due both to the distinct possibility of cosmic ray exposure prior to deposition and to erosional processes modifying both the boulders and the landform after deposition. The susceptibility of any particular boulder to erosion, as well as the time at which it was exposed on the landform surface, will determine the concentration of cosmogenic isotopes. For example, *Jull et al.* [1992] find lower (< 10% to 60%) than expected abundances of cosmogenic ^{14}C in four out of 11 samples used by *Dorn et al.* [1991] for rock varnish and ^{36}Cl dating of moraines in Hawaii and suggest that the abundance of longer-lived isotopes is in part controlled by erosion rather than exposure age. Prior exposure of clasts on alluvial fans and beach ridges is suggested by the ^{26}Al and ^{10}Be data of *Nishiizumi et al.* [1993] and by the depth/production rate calculations for ^{36}Cl of *Bierman* [1993].

In contrast to the broad exposure age distributions of boulders on some morainal deposits, the small dispersion of lava flow ages reported by *Cerling* [1990] and *Zreda et al.* [1993] and of latest Wisconsinan morainal boulder ages reported by *Evenson and Gosse* [1993] suggests that in some geomorphic situations exposure ages may provide relatively precise (+/-10%) estimates of landform age; the accuracy of such data is more difficult to assess but depends heavily on the accuracy of production rates used to interpret isotope abundances.

Model Erosion Rates

Understanding rates of denudation on a variety of time and length scales is important for quantifying the relationship between tectonics and topography. In order to model long-term behavior of continents and mountain ranges, rates of denudation need to be estimated at large spatial (>100 km^2) and temporal (>100 kyr) scales. Conversely, to understand better the evolution of tectonically related features such as fault scarps and escarpments, erosion rates need to be determined at significantly shorter time and length scales.

Rates of denudation have been calculated by a variety of means including sediment accumulation in geologic and constructed reservoirs, measurement of sediment transport rates, monitoring of surface processes and determination of mineral cooling ages (Figure 2). A large number of these studies, many of which reflect either historical or Holocene rates of denudation, have been compiled by *Saunders and Young* [1983]. Only a few of the rates compiled by *Saunders and Young* [1983] reflect directly the erosion of subaerially exposed bedrock surfaces. Conversely, existing cosmogenic erosion rate studies have been directed specifically toward understanding the rate at which exposed bedrock surfaces erode.

So far, cosmogenic isotopes have been used to estimate erosion rates in relatively few geologic and geomorphic settings (Figure 2). In some cases, sufficient samples have been collected to provide an estimate of the spatial variability of isotope abundance; in other cases, only a single sample was collected. For example, *Hampel et al.* [1975] measured the abundance of muon-produced ^{26}Al using counting techniques and estimated an erosion rate of about 10 m/m.y. on chert at Kutami Terrace, Japan. *Kubik* [1984] measured ^{36}Cl produced by muon interactions and estimated an erosion rate of 17 m/m.y. for limestone collected from a quarry near Regensburg, Germany, and *Nishiizumi et al.* [1986] used ^{10}Be and ^{26}Al to calculate erosion rates of 28 and 17 m/m.y. for samples from the Anza Borrego desert. In each of these cases, the measurements were made primarily to demonstrate the viability of the cosmogenic technique and so the geomorphic setting from which the samples were collected is considered only briefly, if at all.

Several workers have measured erosion rates on lava flows. *Sarda et al.* [1993] use the discrepancy between ^{21}Ne abundance and a 62.5 kyr K/Ar eruption age to estimate an erosion rate of 3.5 m/m.y. for basalt in the wet, tropical climate of Reunion; their

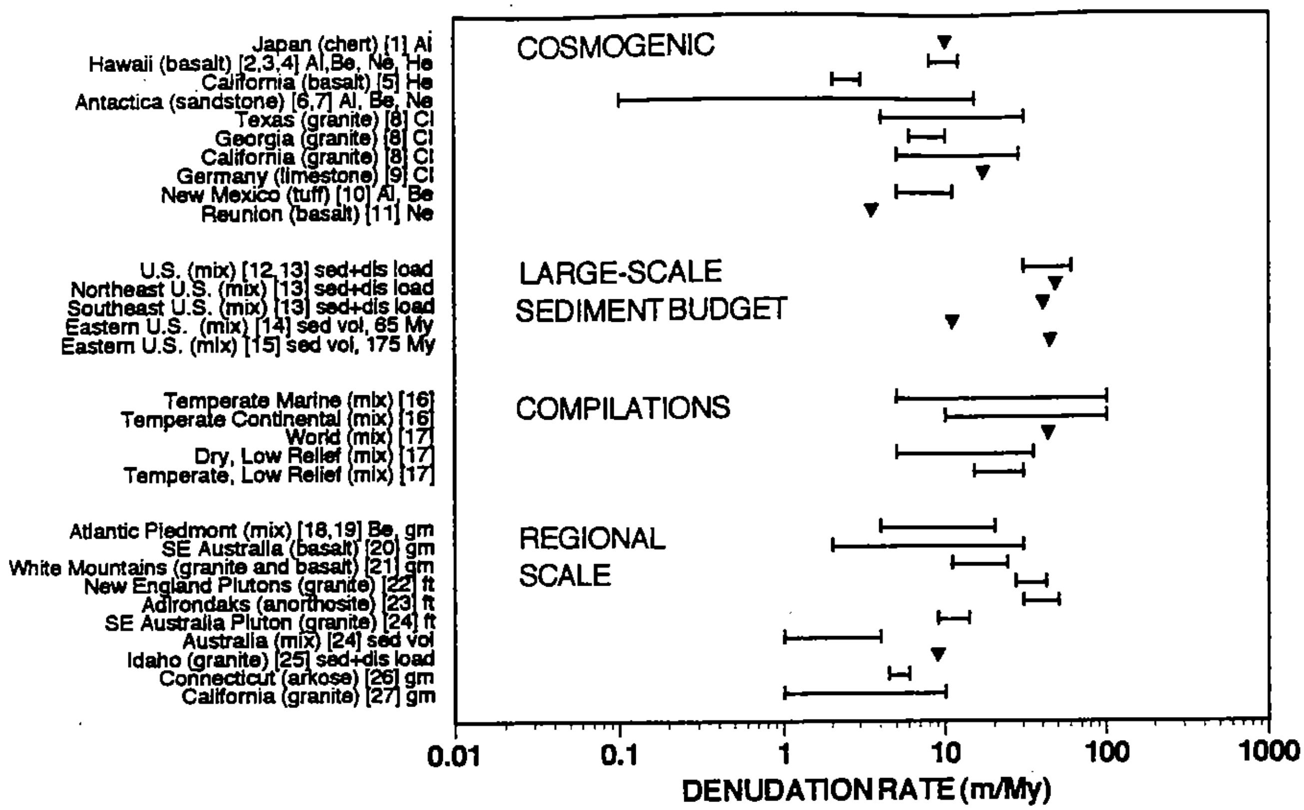

Figure 2. Denudation rates determined cosmogenically and by other means. Rock types are given in parentheses, studies are referenced by numbers in brackets and the method used to estimate denudation rate is given last. For cosmogenic methods, measured elements are indicated, for others: sed+dis load is measured load of sediment and solutes, sed vol is measured sediment volume in basin, gm is geomorphic arguments made on the basis of dated and incised or eroded surfaces, ft is fission tracks. 1, *Hampel et al.* [1975]; 2, *Nishiizumi* [1991]; 3, *Kurz* [1986b]; 4, *Craig and Poreda* [1986]; 5, *Cerling* [1990]; 6, *Nishiizumi et al.* [1991a]; 7, *Staudacher and Allegre* [1991]; 8, *Bierman* [1993]; 9, *Kubik* [1984]; 10, *Albrecht et al.* [1993]; 11, *Sarda et al.* [1993]; 12, *Judson* [1968]; 13, *Judson and Ritter* [1964]; 14, *Mathews* [1975]; 15, *Menard* [1961] rate adjusted for opening of Atlantic at 175 Ma; 16, *Saunders and Young* [1983]; 17, *Summerfield* [1991]; 18, *Pavich et al.* [1985]; 19, *Pavich* [1989]; 20, *Young and McDougall* [1993]; 21, *Marchand* [1971]; 22, *Doherty and Lyons* [1980]; 23, *Miller and Lakatos* [1983]; 24, *Bishop* [1985]; 25, *Clayton and Megahan* [1986]; 26, *Matthias* [1967]; 27, *Oberlander* [1972].

calculation is sensitive to the accuracy of both the young K/Ar age and ^{21}Ne production rates. *Craig and Poreda* [1986] used the abundance of ^{3}He and the assumption of a steady state to calculate an erosion rate of 8.5 m/m.y. for a sample collected from the crest of Haleakala on Maui. *Kurz* [1986b] estimates ^{3}He model erosion rates between 8 and 11 m/m.y. for four other Haleakala samples collected from a variety of elevations. *Nishiizumi* [1991] compares ^{3}He, ^{10}Be, ^{21}Ne and ^{26}Al measurements on a single sample, also from Haleakala, and shows that calculated, steady state erosion rates from different isotope systems are in general agreement and vary between 8 and 12 m/m.y.. Cosmogenically determined (^{36}Cl) erosion rates for granitic landforms in the southern United States are similar to those reported from other locations. Median rates of erosion for the Georgia Piedmont, the Llano Uplift (Texas) and the Alabama Hills (California) are 8, 12, and 7 m/m.y., respectively [*Bierman*, 1993].

Cosmogenically determined erosion rates from Antarctica stand in stark contrast to those determined elsewhere in the world. *Brown et al.* [1992] used a series of ^{10}Be and ^{26}Al measurements in a core to estimate an erosion rate < 0.3 m/m.y. for a quartz sandstone surface in the Quartermain Mountains. In the most extensive erosion rate study to date, *Nishiizumi et al.* [1991a] report ^{26}Al and ^{10}Be abundances for 27 samples collected from four locations in Antarctica. They use these abundances to calculate maximum steady-state erosion rates ranging from <0.1 to 15 m/m.y. Most erosion rates they measured were an order of magnitude lower than those reported from other terrestrial locations, between 0.1 and 1 m/m.y., and likely reflect the near-absence of liquid water and its effect on chemical weathering and mechanical disintegration.

The relatively few "cosmogenic" erosion rate measurements that have been made are generally equal to or lower than those estimated by other means. Cosmogenically based erosion rates for granites, limestones, basalts, tuffs, and cherts in a mix of continental and maritime climate regimes generally range from 2 to 30 m/m.y.. The lowest cosmogenic erosion rates outside of Antarctica were measured on basalts collected from arid southeast California (2-3 m/m.y.). For comparison, most long-term, large-scale denudation rates for North America, estimated using sediment volumes, sediment transport and fission tracks, range from 20 to 60 m/m.y. within the range of erosion rates for similar geographic settings compiled by *Saunders and Young* [1983] and *Summerfield* [1991]. It is likely that at least some of these contemporary and large-scale denudation rates reflect anthropogenic sediment contribution and sediment resulting from >15 glacial/deglacial cycles and thus are higher than the average Cenozoic rate.

Erosion rate studies in smaller basins show rates similar to those calculated cosmogenically; for instance *Marchand* [1971] and *Oberlander* [1972] used dated (10 m.y.) and incised basalt flows to calculate, respectively, average denudation rates between 11 and 24 m/m.y. for an arid part of the White Mountains in California and 1-10 m/m.y. for granite in the Mojave desert. These denudation rates are similar to those measured by *Bierman* [1993] using ^{36}Cl in the nearby granitic Alabama Hills. *Dethier et al.* [1988], working in northern New Mexico, measured denudation rates from 4 m/My (tuff, basalt) to >10 m/m.y. (weakly lithified sandstone). Cosmogenically determined erosion rates for granitic outcrops in the Georgia Piedmont range from 6 to 10 m/m.y. [*Bierman*, 1993] and are similar to those determined by several independent techniques, 4 and 20 m/m.y. [*Pavich et al.*, 1985; *Pavich*, 1989]. Other comparisons are more tenuous. For instance, cosmogenically deduced rates of basalt erosion on Maui appear to be about 10 m/m.y. whereas two measurements from semiarid Owens Valley and one from wet, tropical Reunion would suggest slightly lower rates of several meters per million years. These cosmogenically determined rates fall within the range of erosion rates (2-30 m/m.y.) determined from incision of dated basalt flows in southeast Australia [*Young and McDougall*, 1993].

Cosmogenically determined (point) rates of denudation (2-30 m/m.y.) are similar to those determined by incision of dated surfaces, by tombstone measurements, and by sediment budgets in small to moderate size catchments. However, these rates are, in general, lower than rates determined by fission tracks and continental scale sediment budgets, including those based on present-day sediment loadings and long-term depositional volumes with the caveat that the assumption of equilibrium and the storage of sediment preclude rigorous interpretation of many of these data. From the limited data (cosmogenic and otherwise) that are available, it appears that rates of bedrock erosion in many climatic settings are grossly similar, with the exception of Antarctica, despite differences in climate and lithology. This interpretation is consistent with that of *Ahnert* [1970], who suggested that local relief, rather than precipitation, is the strongest control on sediment yield (denudation rate).

On the basis of the few data available, any explanation of the observed differences can only be speculative. Inadvertent sampling bias may account for what appear to be relatively low rates of erosion calculated from cosmogenic isotope measurements. So far, samples have been collected only from relatively resistant lithologies such as basalt, granite and sandstone. In addition, it appears that samples have generally been collected from outcrops standing above the surrounding landscape. These protrusions of bedrock may erode more slowly because they shed water rapidly thus reducing the efficacy of chemical weathering. Since most cosmogenic samples have been collected to determine landform age rather than erosion rate, there is presumably a bias toward samples collected from gently sloping interfluves rather than steep, and presumably more rapidly eroding, hillslopes. Unfortunately, in many papers, the geomorphic setting of specific samples is described only briefly, if at all.

Geomorphic Process Studies

There are a variety of other ways in which cosmogenic isotope abundances could be used to understand better geomorphic processes. For example, *Wells et al.* [1991] used cosmogenic ^{3}He and ^{21}Ne to support a model for the evolution of desert pavements. They sampled basalt from an exposed flow and basalt clasts which were derived from the same flow but which are now contained in an overlying desert pavement. They found similar isotope abundances in the both sample populations, supporting their model of pavement evolution in which clasts float on an underlying silt-rich horizon [*McFadden et al.*, 1987]. Their application is particularly robust in that the veracity of their findings does not depend on an accurate knowledge of production rates or their variation through time.

From a deep road cut, *Nishiizumi et al.* [1989] sampled the Sherwin Till below the Bishop Tuff (0.73 m.y.) in an attempt to date the till. They conclude that the exposure history of quartz grains in the uppermost portion of the till is complex and not amenable to simple interpretation. However, later, *Nishiizumi et al.* [1993] used a model of steady aggradation to interpret the measured abundance of ^{10}Be and ^{26}Al in quartz grains separated from the till. Their interpretive model appears not to take into account the thickness of the overlying ice, deformation of till below the glacier or the variability of initial isotope abundance in the material that makes up the till. On the basis of what appears to be a geomorphically unrealistic interpretive model, they calculate extremely low rates of till deposition (1 to 3 x 10^{3} cm yr^{-1}).

Conclusions

It is clear that the use of in situ produced cosmogenic isotopes holds promise for defining better the rates of geomorphic processes. There are, however, limitations to the precision, accuracy and most importantly geomorphic meaning of model erosion rates and exposure ages. Currently, analytic and production rate uncertainties in many cases overshadow those related to geologic processes. However, as the study of cosmogenic isotopes continues, these technical uncertainties will likely decrease to the point at which, in most studies, the geologic history of the sample becomes the dominant uncertainty.

The analysis of multiple isotopes for each sample is a viable means by which to constrain the uncertainty of exposure age and erosion rate estimates. Each isotope has particular characteristics which can bias abundance estimates. For instance, atmospheric ^{10}Be can contaminate measurements of the in situ produced isotope, ^{21}Ne and ^{36}Cl can be produced radiogenically in significant quantities, and ^{3}He may not be well retained in some minerals, especially in warm climates. Furthermore, as measurement precision improves and production rates and ratios become better known, measurements of multiple isotopes in single samples will reveal more about the sample's geologic history [*Graf et al.*, 1991; *Lal*, 1991; *Nishiizumi et al.*, 1991a; 1993].

If cosmogenic data are to be geomorphically meaningful, then the assumptions on which interpretive models are based must not only be understood, they must also be valid. For example, apparent correlation between cosmogenically deduced glacial or alluvial chronologies and those determined by other means will be fallacious if erosion has biased cosmogenic exposure ages, particularly if such bias is not a linear function of surface age. In every study to date, exposure age variability on a single geomorphic surface is significant (usually >10%) and in most cases greater than expected from analytical uncertainty. Reliably correlating the age of a geomorphic surface, on which a range of cosmogenic exposure ages has been determined, with other temporal records such as ice or deep-sea sediment cores remains a significant and as yet unresolved problem [*Brook et al.*, 1993]. It is likely that in most cases the mean exposure age for a particular geomorphic surface is controlled, at least in part, by spatially and temporally variable rates of weathering and erosion [*Bierman and Gillespie*, 1991; 1992; *Jull et al.*, 1992] and thus has only limited chronological significance.

Cosmogenic erosion rate estimates appear to be geomorphically useful, despite uncertainties in isotope production rates and the potentially untenable assumption of steady state denudation, primarily because there are so few other constraints on the rate of bedrock erosion. However, extrapolating erosion rates determined at single points on bedrock outcrops to larger spatial and temporal scales introduces a variety of assumptions. Sampling of outcrops disregards soil-covered portions of the landscape which may be eroding at a different rate. If erosion is episodic and decimeters to meters of rock are removed in each erosion episode, cosmogenically based erosion rates could either overestimate or underestimate long-term rates of denudation. So far, only relatively resistant lithologies have been sampled. In addition, most samples have been collected from tectonically stable areas or areas of low relief and do not reflect the high rates of denudation typically found in alpine regions. Perhaps the apparent discrepancy between cosmogenically deduced rates of erosion and those determined over longer spatial and temporal scales would diminish if such potential sampling bias could be overcome.

One could consider two different cosmogenic approaches to the problem of measuring long-term, basin-scale rates of erosion. A surface-based approach would require collecting samples from a variety of geomorphic environments within the basin, mapping the spatial abundance of each environment and calculating a spatially averaged rate of erosion. Such an approach would be hampered by the difficulty one faces in determining which areas have similar erosion rates, the necessary assumption of steady state and the need to collect and analyze a large number of samples. Conversely, *Bierman and Steig* [1992, in review] suggest that it may be possible to measure the abundance of cosmogenic isotopes in sediment and deduce the erosion rate of the basin supplying these sediments. Their approach requires that outgoing sediments are well mixed, that the basin is in isotopic steady state, and that a stable isotope is measured or erosion rates are greater than several meters per million years.

It appears that the most useful applications of cosmogenic nuclides will consider time and length scales over which model assumptions are reasonably valid and will be backed up by robust geomorphic constraints. At present, scales at which cosmogenic isotopes are most useful appear to be shorter than those of interest to tectonicists modeling the development and decay of orogenic belts. Dating of tectonically offset landscape features could help constrain rates of faulting and surface uplift. However, the depth of boulder weathering and amount of landform erosion are both difficult to quantify and thus it is difficult to correct accurately for their effects on cosmogenic age estimates. Nevertheless, in some cases, cosmogenic isotope abundances can be used to provide minimum limiting ages for geomorphic surfaces; in other cases, such abundances can be used to constrain the maximum rate at which a specific bedrock outcrop has eroded. Considering current uncertainties in isotope production rates and the validity of model assumptions, only carefully chosen geomorphic problems will yield data that allow meaningful and rigorous interpretation.

Acknowledgments. I wish to thank the following individuals for discussions or reviews which are reflected in the content of this paper: A. Gillespie, E. Steig, F. Phillips, D. Elmore, L. Dep, M. Caffee, B. Hudson, and B. Nelson. Reviews of this paper by S. Wells, D. Merritts, D. Dethier, M. Pavich, and an anonymous individual significantly improved the original manuscript. Research supported by NSF grants EAR-9004252 and EAR-9219487 and USGS grant 14-08-001-G1783. I thank A. Church for preparing camera ready copy.

References

Ahnert, F., Functional relationships between denudation, relief, and uplift in large mid-latitude drainage basins, *Am. J. Sci., 268*, 243-263, 1970.

Albrecht, A., G. Herzog, J. Klein, B. Dezfouly-Arjomandy, and F. Goff, Quaternary erosion and cosmic-ray-exposure history derived from ^{10}Be and ^{26}Al produced in situ --An example from Pajarito plateau, Valles caldera region, *Geology, 21*, 551-554, 1993.

Andrews, J. N., J.-C. Fontes, J.-L. Michelot, and D. Elmore, In situ ^{36}Cl production and groundwater evolution in crystalline rocks at Stripa, *Earth Planet. Sci. Lett., 77*, 49-58, 1986.

Andrews, J. N., S. N. Davis, J. Fabryka-Martin, J. C. Fontes, B. Lehmann, H. H. Loosli, J.-L. Michelot, H. Moser, B. Smith and M. Wolf, The in situ production of radioisotopes in rock matrices with particular reference to the Stripa granite, *Geochim. Cosmochim. Acta, 53*, 1803-1815, 1989.

Anthony, E., and J. Poths, ^{3}He surface exposure dating and its implications for magma evolution in the Potrillo volcanic field, Rio Grande Rift, New Mexico, USA, *Geochim. Cosmochim. Acta, 56*, 4105-4108, 1992.

Bard, E., B. Hamelin, R. G. Fairbanks, and A. Zindler, Calibration of the ^{14}C timescale over the past 30,000 years using mass spectrometric U-Th ages from Barbados corals, *Nature, 345*, 405-410, 1990.

Bentley, H. W., F. M. Phillips, S. N. Davis, M. A. Habermehl, P. Airey, G. E. Calf, D. Elmore, H. E. Gove and T. Torgersen, Chlorine 36 dating of very old ground water, 1, The great artesian basin, Australia, *Water Resour. Res. 22*, 1991-2001, 1986.

Bierman, P., Cosmogenic isotopes and the evolution of granitic landforms, doctoral thesis, Univ. of Wash., Seattle, 1993.

Bierman, P. R., and A. R. Gillespie, Range fires: A significant factor in exposure-age determination and geomorphic surface evolution, *Geology, 19*, 641-644, 1991.

Bierman, P. R., and A. R. Gillespie, Reply to comment on Range fires: A significant factor in exposure-age determination and geomorphic surface evolution, *Geology, 20*, 283-285, 1992.

Bierman, P. R., and E. Steig, Using cosmogenic isotopes to measure basin-scale rates of erosion., *Geol. Soc. Am. Abstr. Programs, 24*, A122, 1992.

Bilokon, H., G. C. Castagnoli, A. Castellina, B. D. Piazzoli, G. Mannocchi, E. Meroni, P. Picchi, and S. Vernetto, Flux of vertical negative muons stopping at depths 0.35-1000 hg/cm^2, *J. Geophys. Res., 94*, 12,145-12,152, 1989.

Birkeland, P. W., *Soils and Geomorphology*, 372 pp., Oxford University Press, New York, 1984.

Bishop, P., Southeast Australian late Mesozoic and Cenozoic denudation rates: A test for late Tertiary increases in continental denudation, *Geology, 13*, 479-482, 1985.

Blackwelder, E., Fire as an agent in rock weathering, *J. Geol., 35*, 134-140, 1927.

Brook, E. J., and M. D. Kurz, Surface-exposure chronology using in situ cosmogenic ^{3}He in Antarctic quartz sandstone boulders, *Quat. Res., 39*, 1-10, 1993.

Brook, E. J., M. D. Kurz, G. H. Denton, and R. P. J. Ackert, Chronology of Taylor Glacier advances in Arena Valley, Antarctica using in situ cosmogenic ^{3}He and ^{10}Be, *Quat. Res., 39*, 11-23, 1993.

Brown, E. T., J. M. Edmond, G. M. Raisbeck, F. Yiou, M. D. Kurz, and E. J. Brook, Examination of surface exposure ages of Antarctic moraines using in situ produced ^{10}Be and ^{26}Al, *Geochim. Cosmochim. Acta, 55*, 2269-2283, 1991.

Brown, E. T., E. J. Brook, G. M. Raisbeck, F. Yiou, and M. D. Kurz, Effective attenuation of cosmic rays producing ^{10}Be and ^{26}Al in Quartz: Implications for exposure dating, Geophys. *Res. Lett., 19*, 369-372, 1992.

Bursik, M. I., and A. R. Gillespie, Late Pleistocene glaciation of Mono Basin, California, *Quat. Res.*, *39*, 24-35, 1993.

Cerling, T. E., Dating geomorphic surfaces using cosmogenic ^{3}He, *Quat. Res.*, *33*, 148-156, 1990.

Charalambus, S., Nuclear transmututation by negative stopped muons and the activity induced by cosmic-ray muons, *Nucl. Phys. A*, *166*, 145-161, 1971.

Clayton, J. L., and W. F. Megahan, Erosional and chemical denudation rates in the southwestern Idaho batholith, *Earth Surf. Processes*, *11*, 389-400, 1986.

Craig, H., and R. J. Poreda, Cosmogenic ^{3}He in terrestrial rocks: The summit lavas of Maui, *Proc. Natl. Acad. Sci. U.S.A.*, *83*, 1970-1974, 1986.

Davis, R., and O. A. Schaeffer, Chlorine-36 in nature, *Ann. N. Y. Acad. Sci.*, *62*, 105-122, 1955.

Dethier, D. P., C. D. Harrington and M. J. Aldrich, Late Cenozoic rates of erosion in the western Espanola Basin, New Mexico: Evidence from geologic dating of erosion surfaces, *Geol. Soc. Am. Bull.*, *100*, 928-937, 1988.

Doherty, J. T., and J. B. Lyons, Mesozoic erosion rates in northern New England. *Geol. Soc. Am. Bull.*, *91*, 16-20, 1980.

Dorn, R. I., F. Phillips, M. Zreda, E. W. Wolfe, A. J. T. Jull, D. J. Donahue, P. W. Kubik, and P. Sharma, Glacial chronology, *Natl. Geogr. Res. Explor.* *7*, 456-472, 1991.

Elmore, D., and F. Phillips, Accelerator mass spectrometry for measurement of long-lived radioisotopes, *Science*, *236*, 543-550, 1987.

Evenson, E., and J. Gosse, Application of in situ produced cosmogenic nuclide exposure ages to reconstruct glacial histories at the Pinedale type locality, Wyoming, *Geol. Soc. Am. Abstr. Programs*, *25*, A308, 1993.

Fabryka-Martin, J. T., Production of radionuclides in the earth and their hydrogeologic significance, with emphasis on chlorine-36 and iodine-129, Ph.D. thesis, Univ. of Ariz., Tucson, 1988.

Feige, G., B. G. Oltman, and J. Kastner, Production rates of neutrons in soils due to natural radioactivity, *J. Geophys. Res.*, *73*, 3135-3142, 1968.

Gillespie, A. R., and P. R. Bierman, Theoretical precision estimates for exposure-age and erosion-rate data from analysis of cosmogenic isotopes, *Geol. Soc. Am. Abstr. Programs*, *23*, A409, 1991.

Graf, T., C. Kohl, K. Marti, and K. Nishiizumi, Cosmic-ray produced Ne in Antarctic rocks, *Geophys. Res. Lett.*, *18*, 203-206, 1991.

Hampel, W., J. Takagi, K. Sakamoto, and S. Tanaka, Measurement of muon-induced ^{26}Al in terrestrial silicate rock, *J. Geophys. Res.*, *80*, 3757-3760, 1975.

Hudson, B., M. Caffee, B. Beirger, B. Ruiz, C. P. Kohl, and K. Nishiizumi, Production rate and retention properties of cosmogenic ^{3}He and ^{21}Ne in quartz, *Eos Trans. AGU*, *72* (44), Fall meeting suppl., 575, 1991.

Judson, S., Erosion of the land or what's happening to our continents, *Am. Sci.*, *56*, 356-374, 1968.

Judson, S., and D. Ritter, Rates of regional denudation in the United States, *J. Geophys. Res.*, *69*, 3395-3401, 1964.

Jull, A. J., A. Wilson, G. Burr, L. J. Toolin, and D. J. Donahue, Measurements of cosmogenic ^{14}C produced by spallation in high-altitude rocks, *Radiocarbon*, *34*, 737-744, 1992.

Klein, J., R. Giegengack, R. Middleton, P. Sharma, J. R. Underwood, and R. A. Weeks, Revealing histories of exposure using in situ produced ^{26}Al and ^{10}Be in Libyan desert glass, *Radiocarbon*, *28*, 547-555, 1986.

Kohl, C. P., and K. Nishiizumi, Chemical isolation of quartz for measurement of *in-situ*-produced cosmogenic nuclides, *Geochim. Cosmochim. Acta*, *56*, 3583-3587, 1992.

Kubik, P., Accelerator mass spectrometry of ^{36}Cl in limestone and some paleontological samples using completely stripped ions, *Nucl. Instrum. Method Phys. Res., Section, Sect. B*, *5*, 326-330, 1984.

Kurz, M. D., Cosmogenic helium in a terrestrial igneous rock, *Nature*, *320*, 435-439, 1986a.

Kurz, M. D., In situ production of terrestrial cosmogenic helium and some applications to geochronology, *Geochim. Cosmochim Acta*, *50*, 2855-2862, 1986b.

Kurz, M. D., D. Colodner, T. W. Trull, R. Moore, and K. O'Brien, Cosmic ray exposure dating with in situ produced cosmogenic ^{3}He: Results from young Hawaiian lava flows, *Earth Planet. Sci. Lett.*, *97*, 177-189, 1990.

Lal, D., On the study of continental erosion rates and cycles using cosmogenic ^{10}Be and ^{26}Al and other isotopes, in *Dating Young Sediments*, edited by A. J. Hurford, E. Jager and J. A. M. Ten Cate, pp. 285-298, CCOP Technical Secretariat, Bangkok, 1986.

Lal, D., In situ-produced cosmogenic isotopes in terrestrial rocks, *Annu. Rev. Earth Planet. Sci.*, *16*, 355-388, 1988.

Lal, D., Cosmic ray labeling of erosion surfaces, In situ production rates and erosion models, *Earth Planet. Sci. Lett.*, *104*, 424-439, 1991.

Lal, D., and J. R. Arnold, Tracing quartz through the environment, *Proc. Indian Acad. Sci.*, *94*, 1-5, 1985.

Lal, D., and B. Peters, Cosmic ray produced radioactivity in the Earth, in *Handbuch der Physik*, edited by K. Sitte, pp. 551-612, Springer-Verlag, New York, 1967.

Leavy, B., Surface-exposure dating of young volcanic rocks using the in situ buildup of cosmogenic isotopes, Ph. D. thesis, N. M. Inst. of Min. and Technol., Socorro, 1987.

Leavy, B. D., F. M. Phillips, D. Elmore, P. W. Kubik, and E. Gladney, Measurement of cosmogenic ^{36}Cl/Cl in young volcanic rocks: An application of accelerator mass spectrometry in geochronology, *Nucl. Instrum. Methods Phys. Res., Sect. B*, *29*, 246-250, 1987.

Lingenfelter, R. E., Production of carbon 14 by cosmic-ray neutrons, *Rev. Geophys.*, *1*, 35-55, 1963.

Marchand, D. E., Rates and modes of denudation, White Mountains, eastern California, *Am. J. Sci.*, *270*, 109-135, 1971.

Mathews, W. H., Cenozoic erosion and erosion surfaces of eastern North America, *Am. J. Sci.*, *275*, 818-824, 1975.

Matthias, G. F., Weathering rates of Portland arkose tombstones, *J. Geol. Educ.*, *15*, 140-144, 1967.

Mazaud, A., C. Laj, E. Bard, M. Arnold, and E. Tric, Geomagnetic field control of ^{14}C production over the last 80 ky: Implications for the radiocarbon time scale, *Geophys. Res. Lett.*, *18*, 1885-1888, 1991.

McElhinny, M. W., and W. E. Senanayake, Variations in the geomagnetic dipole 1, The past 50,000 years, *J. Geomagn. Geoelectr.*, *34*, 39-51, 1982.

McFadden, L. D., S. G. Wells and M. J. Jercinovich, Influences of eolian and pedogenic processes on the origin and evolution of desert pavements, *Geology*, *15*, 504-508, 1987.

McFadden, P. L., and M. W. McElhinny, Variations in the geomagnetic dipole 2: statistical analysis of VDMs for the past 5 million years, *J. Geomagn. Geoelectr.* *34*, 163-189, 1982.

Menard, H. W., Some rates of regional erosion, *J. Geol.*, *69*, 154-161, 1961.

Middleton, R., and J. Klein, ^{26}Al: Measurement and applications, *Philos. Trans. R. Soc. London, Ser. A*, *323*, 121-143, 1987.

Miller, D. S., and S. Lakatos, Uplift rate of Adirondack anorthosite measured by fission-track analysis of apatite, *Geology*, *11*, 284-286, 1983.

Nishiizumi, K., Cosmogenic 10-Be, 26-Al, and 3-He in olivine from Maui lavas, *Earth Planet. Sci. Lett.*, *98*, 263-266, 1991.

Nishiizumi, K., D. Lal, J. Klein, R. Middleton and J. R. Arnold, Production of ^{10}Be and ^{26}Al by cosmic rays in terrestrial quartz in situ and implications for erosion rates, *Nature*, *319*, 134-136, 1986.

Nishiizumi, K., E. Winterer, C. P. Kohl, J. Klein, R. Middleton, D. Lal, and J. R. Arnold, Cosmic ray production rates of ^{10}Be and ^{26}Al in quartz from glacially polished rocks, *J. Geophys. Res.*, *94*, 17907-17915, 1989.

Nishiizumi, K., C. P. Kohl, J. R. Arnold, J. Klein, D. Fink, and R. Middleton, Cosmic ray produced ^{10}Be and ^{26}Al in Antarctic rocks: Exposure and erosion history, *Earth Planet. Sci. Lett.*, *104*, 440-454, 1991a.

Nishiizumi, K., C. P. Kohl, E. M. Shoemaker, J. R. Arnold, J. Klein, D. Fink, and R. Middleton, In situ ^{10}Be-^{26}Al exposure ages at Meteor Crater, Arizona, *Geochim. Cosmochim. Acta*, *55*, 2699-2703, 1991b.

Nishiizumi, K., C. P. Kohl, J. R. Arnold, R. I. Dorn, J. Klein, D. Fink, R. Middleton, and D. Lal, Role of in situ cosmogenic nuclides ^{10}Be and

^{26}Al in the study of diverse geomorphic processes, *Earth Surf. Processes, 18*, 407-425, 1993.

O'Brien, K., H. A. Sandmeier, G. E. Hansen, and J. E. Campbell, Cosmic ray induced neutron background sources and fluxes for geometries of air over water, ground, iron, and aluminum, *J. Geophys. Res., 83*, 114-120, 1978.

Oberlander, T., Morphogenesis of granitic boulder slopes in the Mojave desert, *J. Geol., 80*, 1-20, 1972.

Ozima, M., and F. A. Podosek, *Noble Gas Geochemistry*, Cambridge University Press, New York, 1983.

Pavich, M. J., Regolith residence time and the concept of surface age of the piedmont "peneplain", *Geomorphology, 2*, 181-196, 1989.

Pavich, M. J., L. Brown, J. N. Valette-Silver, J. Klein, and R. Middleton, ^{10}Be analysis of a Quaternary weathering profile in the Virginia Piedmont: *Geology, 13*, 39-41, 1985.

Phillips, F. M., M. G.. Zreda, S. S. Smith, D. Elmore, P. W. Kubik and P. Sharma, Cosmogenic chlorine-36 chronology for glacial deposits at Bloody Canyon, eastern Sierra Nevada, *Science, 248*, 1529-1532, 1990.

Phillips, F. M., M. G.· Zreda, S. S. Smith, D. Elmore, P. W. Kubik, R. I. Dorn and D. J. Roddy, Age and geomorphic history of Meteor Crater, Arizona from cosmogenic ^{36}Cl and ^{14}C in rock varnish, *Geochim. Cosmochim. Acta, 55*, 2695-2698, 1991.

Phillips, F., M. G. Zreda, and D. Elmore, A recalibration of cosmogenic chlorine-36 production rates using lava flow samples from the western Snake River Plain volcanic field, *Geol. Soc. Am. Abstr. Programs, 25*, A373, 1993.

Pomerantz, M. A., and S. P. Agarwal, Spatial distribution of cosmic ray intensity and geomagnetic theory, *Philos. Mag., 7*, 1503-1511, 1962.

Poreda, R., and T. Cerling, Cosmogenic neon in recent lavas from the western United States, *Geophys. Res. Lett., 19*, 1863-1866, 1992.

Raisbeck, G. M., and F. Yiou, Dating by cosmogenic isotopes ^{10}Be, ^{26}Al, ^{41}Ca, in *Nuclear Methods of Dating*, edited by E. Roth and B. Poty, pp. 353-378, Kluwer, Norwell, Mass., 1990.

Raisbeck, G. M., F. Yiou, J. Klein and R. Middleton, Accelerator mass spectrometer measurement of cosmogenic ^{26}Al in terrestrial and extraterrestrial matter, *Nature, 301*, 690-692, 1983.

Raisbeck, G. M., F. Yiou and D. Bourles, Evidence for an increase in cosmogenic ^{10}Be during a geomagnetic reversal, *Nature, 315*, 315-317., 1985.

Reedy, R. C., J. R. Arnold and D. Lal, Cosmic ray record in solar system matter, *Science, 219*, 127-135, 1983.

Rose, D. C., K. B. Fenton, J. Katzman and J. A. Simpson, Latitude effect of the cosmic ray nucleon and meson components at sea level from the Arctic to the Antarctic, *Can. J. Phys., 34*, 968-984, 1956.

Sarda, P., T. Staudacher, C. Allegre, and A. Lecomte, Cosmogenic neon and helium at Reunion: Measurement of erosion rate, *Earth Planet. Sci. Lett., 119*, 405-417, 1993.

Saunders, I., and A. Young, Rates of surface processes on slopes, slope retreat, and denudation, *Earth Surf. Processes, 8*, 473-501, 1983.

Sharma, P., and R. Middleton, Radiogenic production of ^{10}Be and ^{26}Al in uranium and thorium ores: Implications for studying terrestrial samples containing low levels of ^{10}Be and ^{26}Al, *Geochim. Cosmochim. Acta, 53*, 709-716, 1989.

Staudacher, T., and C. J. Allegre, Cosmogenic neon in ultramafic nodules from Asia and in quartzite from Antarctica, *Earth Planet. Sci. Lett., 106*, 87-102, 1991.

Staudacher, T., and C. J. Allegre, Ages of the second caldera of Piton de la Fournaise volcano (Reunion) determined by cosmic ray produced ^{3}He and ^{21}Ne, *Earth Planet. Sci. Lett., 119*, 395-404, 1993a.

Staudacher, T., and C. J. Allegre, The cosmic ray produced ^{3}He/^{21}Ne ratio in ultramafic rocks, *Geophys. Res. Lett., 20*, 1075-1078, 1993b.

Summerfield, M., *Global Geomorphology*, pp. 537, Longman, New York, 1991.

Swanson, T., M. Caffee, R. Finkel, L. Harris, J. Southon, M. Zreda, and F. Phillips, Establishment of new production parameters for chlorine-36 dating based on the deglaciation history of Whidbey Island, Washington, *Geol. Soc. Am. Abstr. Programs, 25*, A-461, 1993.

Swanson, T. W., P. Sharma, F. M. Phillips, and M. G. Zreda, Determination of ^{36}Cl production rates from the deglaciation of Whidbey Island, Washington, paper presented at 203d American Chemical Society Meeting, San Francisco, Calif., 1992.

Tieh, T. T., and E. B. Ledger, Fission track study of Uranium in two granites of central Texas, *Contrib. Mineral. Petrol., 76*, 12-16, 1981.

Tric, E., J. P. Valet, P. Tucholka, M. Paterne, L. Labeyrie, F. Guichard, L. Tauxe and M. Fontugne, Paleointensity of the geomagnetic field during the last 80,000 years, *J. Geophys. Res., 97*, 9337-9351, 1992.

Turner, J., Location of uranium and thorium in granites--Radiogenic influence on cosmogenic isotope abundance, *Geol. Soc. Am. Abstr. Programs, 25*, A-89, 1993.

Wells, S. G., L. D. McFadden, and C. T. Olinger, Use of cosmogenic ^{3}He and ^{21}Ne to understand desert pavement formation, *Geol. Soc. Am. Abstr. Programs, 23*, A207, 1991.

Whitney, J. W., and C. D. Harrington, Relict colluvial boulder deposits as paleoclimatic indicators in the Yucca Mountain region, southern Nevada, *Geol. Soc. Am. Bull., 105*, 1008-1018, 1993.

Yokoyama, Y., J. Reyss, and F. Guichard, Production of radionuclides by cosmic rays at mountain altitudes, *Earth Planet. Sci. Lett., 36*, 44-50, 1977.

Young, R., and I. McDougall, Long-term landscape evolution, Early Miocene and modern rivers in southern New South Wales, Australia, *J. Geol., 101*, 35-49, 1993.

Zito, R., D. J. Donahue, S. N. Davis, H. W. Bentley and P. Fritz, Possible subsurface production of carbon-14, *Geophys. Res. Lett., 7*, 235-238, 1980.

Zreda, M. G., F. M. Phillips, D. Elmore, P. W. Kubik, P. Sharma and R. I. Dorn, Cosmogenic chlorine-36 production rates in terrestrial rocks, *Earth Planet. Sci. Lett., 105*, 94-109, 1991.

Zreda, M., F. M. Phillips, P. W. Kubik, P. Sharma and D. Elmore, Cosmogenic ^{36}Cl dating of a young basaltic eruption complex, Lathrop Wells, Nevada, *Geology, 21*, 57-60, 1993.

P.R. Bierman, Department of Geology, University of Vermont, Burlington, VT 05404. (e-mail: pbierman@moose.uvm.edu)

(Received December 7, 1993; revised February 3, 1994; accepted February 14, 1994.)