Geol. 655 Isotope Geochemistry

lecture 13 Spring 2003

COSMOGENIC NUCLIDES II

IN-SITU-PRODUCED COSMOGENIC NUCLIDES

A few cosmic rays and secondary particles manage to pass entirely through the atmosphere where they interact with rock at the surface of the Earth. These interactions are also capable of producing ^{36}Cl, ^{26}Al, and ^{10}Be as well as many other stable and unstable nuclei. Nuclides most likely to be of geological use are listed in Table 1 along with their production mechanisms. Because the atmosphere is a very effective cosmic ray shield, these cosmogenic nuclides are rare and difficult to detect, so most of the nuclides of interest are unstable ones that otherwise do not occur in the Earth. However, stable cosmogenic nuclides can also provide useful geological information. To be of use, the cosmogenic production must be large relative compared to the background abundance of such nuclides. Thus only ^{3}He and ^{21}Ne have been studied to date. In the following examples, we will see how the cosmogenic production of a rare stable nuclide, ^{3}He, can be used to estimate erosion rates, and how ^{36}Cl can be used to determine the time material has been exposed to cosmic rays.

TABLE 13.1. ISOTOPES WITH APPRECIABLE PRODUCTION RATES IN TERRESTRIAL ROCKS.

Isotope	Half-life (10^{-15})	Thermal neutrons Target	Reaction	Capture of μ⁻ Target	Low-energy □ particles Target	Reaction
^{3}He	stable	^{6}Li	(n,□)		^{11}B	(□,t)
^{10}Be	1.6 □ 10^{6}	—	—	^{10}B,C,N,O	^{7}Li	(□,p)
^{14}C	5730	^{14}N,^{17}O	(n,p),(n,□)	N,O	^{11}B	(□,p)
^{21}Ne, ^{22}Ne	stable	—	—	Na,Mg,Al	^{18}O,^{19}F	(□,n)
^{26}Al	7.1□ 10^{5}	—	—	Si,S	^{23}Na, ^{25}Mg	(□,n),(□,t)
^{36}Cl	3.0□ 10^{5}	^{35}Cl,^{39}K	(n,□),(n,□)	K, Ca, Sc	^{33}S	(□,p)
^{129}I	1.6 □ 10^{5}	^{128}Te	(n,□)	^{130}Te, Ba		

Determining Erosion Rates from Cosmogenic ^{3}He

The penetration of cosmic rays decays exponentially with depth according to:

$$e^{-z\rho/l} \qquad 13.1$$

where z is depth, l is a constant that depends on the nature and energy of the particle and on the material it penetrates, and ρ is the density. For the nucleonic component of cosmic rays, l is approximately 160 g-cm^{2}. For a material, such as a typical rock, having a density of 2.5 g/cc, the ratio ρ/l, which could be referred to as the characteristic penetration depth, is about 64 cm. So at a depth of 64 cm, the cosmic ray flux would be 1/e or 0.36 times the flux at the surface. For the μ(muon)* component, l is about 1000g-cm^{2}, and for □ (neutrinos) l is nearly infinite (because neutrinos interact so weakly with matter). Most of the cosmic ray interactions are with the nucleonic component. The meaning of all this is that cosmogenic nuclides will be produced only on the surface (top meter or two) of a solid body. As in the atmosphere, cosmic ray interactions produce both stable and unstable nuclei. If we consider the case of the production of a stable

* The μ particle belongs to the family of particles known as *leptons*, the most familiar members of which are the electron and positron. Like the electron, it may be positively or negatively charged and has a spin of $^{1}/_{2}$. However, its mass is about 100 MeV, more than 2 orders of magnitude greater than that of the electron, and about an order of magnitude less than the proton. It is produced mainly by decay of pions, which are also leptons and are created by high-energy cosmic ray interactions. Muons are unstable, decaying to electrons and positrons and □$_{u}$ (muonic neutrino) with an average lifetime of 2 □ 10^{-6} sec. Because muons are leptons, they are not affected by the strong force, and hence interact more weakly with matter than the nucleonic component of cosmic rays.

nucleus, the number of stable nuclei produced at the surface of the body over some time t is simply given by:

$$N = P\,t \qquad 13.2$$

where P is the production rate at the surface, which is in turn a function of the cosmic ray flux, depth, elevation, geomagnetic latitude, and reaction cross section. If we know the production rate, we can solve 13.2 for t, the length of time the surface has been exposed to cosmic rays.

Despite being the second most abundant element in the cosmos, He is very rare on Earth because it is too light to be retained — it escapes from the atmosphere readily. Of helium's two isotopes, ^{3}He is some six orders of magnitude less abundant than ^{4}He. This is because ^{4}He is continually produced by α-decay. Hence the Earth's supply of ^{4}He is continually replenished, whereas ^{3}He is not†. In 1987 M. D. Kurz found extraordinarily high ^{3}He/^{4}He ratios in basalts from Hawaii. The origin proved to be cosmogenic. Most of the ^{3}He is produced by spallation, with a minor component produced by ^{6}Li(n,α) and ^{7}Li(μ,α) reactions. Figure 13.1 shows the decrease in cosmogenic ^{3}He with depth in a core from Haleakala (Maui, Hawaii) compared with the predicted decrease for l = 165 g-cm^2. The dashed line show the depth dependence of the μ stopping rate needed to explain the discrepancy between the predicted and observed depth dependence. Ignoring the small contribution from muon interactions, the concentration of ^{3}He as a function of depth, z, and exposure time, t, is given by:

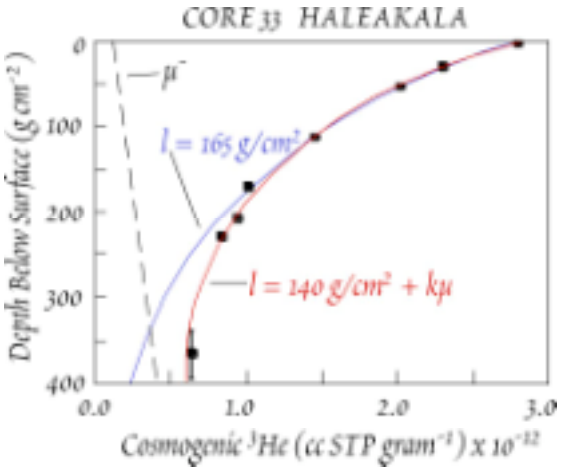


Figure 13.1. Variation of cosmogenic ^{3}He with depth in a core from Haleakala volcano in Hawaii. From Kurz (1986).

$$C(z,t) = \int_0^t P e^{-z\rho/l} dt \qquad 13.3$$

If the depth is not a function of time, this simply integrates to

$$C(z,t) = P e^{-z\rho/l} t \qquad 13.4$$

If erosion occurs, then z will be a function of t. We obtain the simplest relationship between time and depth by assuming the erosion rate is time-independent:

$$z = z_0 - \varepsilon t \qquad 13.5$$

where ε is the erosion rate. Substituting for z in equation 13.3 and integrating, we have

$$C(z,t) = P \frac{l}{\varepsilon\rho} e^{-z_0\rho/l}\left[e^{\varepsilon\rho t/l} - 1\right] \qquad 13.6$$

Substituting $z_0 = z + \varepsilon t$, equ. 13.6 simplifies to:

$$C(z,t) = P \frac{l\, e^{z\rho/l}}{\varepsilon\rho}\left[1 - e^{\varepsilon t}\right] \qquad 13.7$$

Using this procedure, Kurz estimated an erosion rate of 10 m/Ma for Haleakala (Kurz noted that for higher erosion rates, it would be necessary to take account of the muon-produced ^{3}He).

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† This is not strictly true. ^{3}He is produced by ^{6}Li(n,α)^{3}He from spontaneous fission-produced neutrons. However, we might guess that this is a rather improbably reaction. First of all, U is a rare element, and furthermore it rarely fissions. Secondly, Li is a rare element, with typical concentrations of a few 10's of ppm. The probability of a fissogenic neutron finding a ^{6}Li nucleus before it is captured by some other nucleus will therefore not be high. Not surprisingly then, the ^{3}He production rate can be considered insignificant in most situations.

Erosion Rates from Radioactive Cosmogenic Nuclides

For a radioactive nuclide such as ^{26}Al or ^{36}Cl we need to consider its decay as well as its production. The concentration of such nuclide as a function of time and depth is given by:

$$C(z,t) = \frac{P\, e^{-z\rho/\Lambda}}{\lambda + \varepsilon\rho/\Lambda}\left[1 - e^{(-\lambda + \varepsilon\rho/\Lambda)t}\right] \qquad 13.8$$

where λ is the decay constant and t is the age of the rock. If the rock is much older that the half-life of the nuclide (i.e., $\lambda t >> 1$; for ^{36}Cl, for example, this would be the case for a rock > 3 Ma old), then the last term tends to 1. Eventually, production of the nuclide, its decay, and erosion will reach steady-state (assuming cosmic ray flux and erosion rate are time-independent). In this case, the concentration at the surface will be given by:

$$C_0 = \frac{P}{\lambda + \varepsilon\rho/\Lambda} \qquad 13.9$$

Since this equation does not contain a time term, we cannot deduce anything about time in this situation. However, knowing the penetration depth, Λ, and the production rate, we can deduce the erosion rate.

^{36}Cl Dating of Glacial Deposits

Let us next consider the build up of a radioactive nuclide in a rock where the erosion rate can be ignored. For a nuclide being both produced by cosmic ray bombardment and lost by radioactive decay, our basic equation becomes:

$$\frac{dN}{dt} = P - \lambda N \qquad 13.10$$

To obtain the abundance, N, of the radionuclide at some time t, we simply integrate 13.10:

$$N = \frac{P}{\lambda}(1 - e^{-\lambda t}) \qquad 13.11$$

For $t >> \lambda$; i.e., after many half-lives, a steady-state is reached where:

$$N = \frac{P}{\lambda} \qquad 13.12$$

For shorter times, however, we can solve equation 13.11 for t. In this case, t is the time the rock has been exposed to cosmic rays. Since the penetration of cosmic rays is so limited, this is the time the rock has been exposed at the surface of the Earth. This particular problem is of some interest in dating rock varnishes and glacial moraines. All moraines but those from the most recent glaciation will be too old for ^{14}C dating, but virtually the entire Pleistocene glacial history is an appropriate target for dating with ^{26}Al or ^{36}Cl.

Since ^{36}Cl is a fairly heavy nuclide, only a few specific cosmic-ray induced nuclear reactions yield ^{36}Cl. The principle modes of production are thermal neutron capture by ^{35}Cl (the most abundant of chlorine's two stable isotopes), spallation reactions on ^{39}K and ^{40}Ca, and muon capture by ^{40}Ca (Phillips et al., 1986). In effect, this

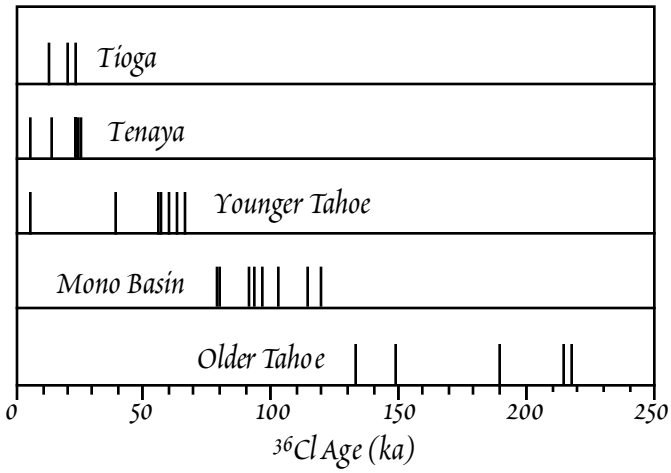


Figure 13.2. ^{36}Cl ages of moraine boulders from Bloody Canyon, eastern Sierra Nevada. (From Phillips et al., 1990)

means the composition of the sample, in particular the concentrations of Cl, K, and Ca, must be known to estimate the production rate. Phillips et al. (1986) showed that the build-up of ^{36}Cl in rocks can be reasonably predicted from these concentrations by determining ^{36}Cl in a series of well-dated lavas and tuffs. In addition to rock composition, it is also necessary to take into consideration 1.) latitude, 2.) elevation, and 3.) non-cosmogenic production of ^{36}Cl. As we saw in the last lecture, spontaneous fission of U and Th will produce neutrons that will result in some production of ^{36}Cl by neutron caption by ^{35}Cl.

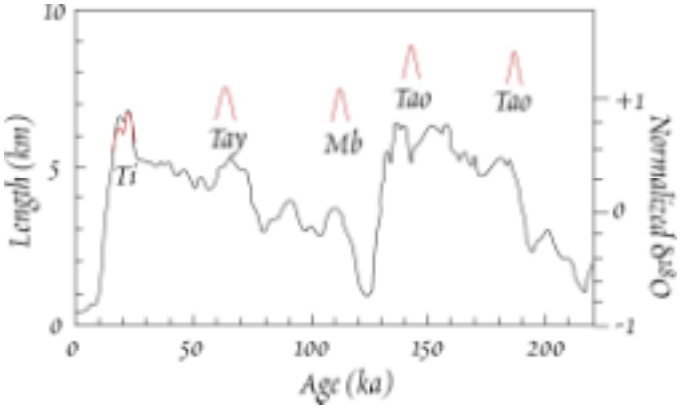


Figure 13.3. Comparison of best estimated ^{36}Cl ages of moraine boulders from Bloody Canyon with the marine ^{18}O record. There is a reasonably good correspondence with the moraine ages and glacial maxima inferred from ^{18}O (from Phillips et al., 1990).

Philips et al. (1990) determined ^{36}Cl ages for boulders taken from a series of moraines in Bloody Canyon of Mono Basin, California. They were careful to sample only boulders from moraine crests as these were most likely to remain above the snow during winter and less likely to have rolled. Their results are shown graphically in Figures 13.2. The youngest moraines correspond to glacial maxima of the most recent glaciation and yield ages in good agreement with ^{14}C dating. Older moraines show considerably more scatter. In additional to analytical errors, factors that might account for the larger scatter include: ^{35}Cl inherited from earlier exposure, preferential leaching of ^{35}Cl, erosion of the rock surface, gradual exposure as a result of erosion of till matrix, and snow cover. Most of these factors will result in the age being too young, so that maximum ages were preferred for the older moraines. The best estimates of moraine ages are compared with the marine O isotope record, which in this case is used as a proxy for global temperature history, in Figure 13.3. Generally, the moraines correspond in time to high δ^{18}O in the oceans, which corresponds to cold temperatures. This is just what we expect: maximum extent of the glaciers occurred during cold climatic episodes.

REFERENCES AND SUGGESTIONS FOR FURTHER READING

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