This paper was peer-reviewed for scientific content.
Pages 1044-1052. In: D.E. Stott, R.H. Mohtar and G.C. Steinhardt (eds). 2001. Sustaining the Global Farm. Selected papers from the 10th International Soil Conservation Organization Meeting held May 24-29, 1999 at Purdue University and the USDA-ARS National Soil Erosion Research Laboratory.

New Nuclear Methods for Studies of Soil Dynamics Utilizing Cosmic Ray Produced Radionuclides

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ABSTRACT

Since soils are primarily "open systems", containing a suite of chemical compounds, which are labile and freely exchange elements and isotopes with the environment, studies of soil dynamics cannot be carried out with the conventional nuclear methods. In this paper we discuss the studies to date, and the promise of three cosmic ray produced (cosmogenic) radionuclides for elucidating key soil dynamics processes.

There are two principal ways in which these radionuclides are produced. The radionuclides, ^{10}Be and ^{32}Si are initially produced in the earth's atmosphere (in nuclear interactions of cosmic rays with nitrogen and oxygen, and argon, respectively), and subsequently brought into soil by rainfall: they serve as useful tracers for determining soil formation and erosion rates. The radionuclides ^{10}Be and ^{14}C are produced not only in the atmosphere but are also produced directly, *in situ*, by nuclear interactions of cosmic rays with oxygen in the mineral grains present in the soil. These nuclides serve as useful tracers for providing reliable information on the rates of erosion of soil over time scales of few thousands to few millions of years.

Our present-day knowledge of the production and dispersion of the cosmogenic nuclides in the atmosphere, hydrosphere and lithosphere, and simple soil weathering and erosion models are discussed. Model soil erosion and soil formation rates are presented based on atmospheric ^{10}Be data. They underscore the value of the cosmogenic tracers for studies of soil dynamics. Since the source strengths of the cosmogenic radionuclides, both of the atmospheric and the *in situ* varieties are fairly well known, they should find applications in wide ranging situations of soil formation and loss through erosion.

INTRODUCTION

An important development in earth sciences in the last half-century was the diverse applications of a wide variety of isotopic methods for studying geophysical/geochemical processes, and for determining their time scales. For process studies, stable isotopes, in particular those belonging to the elements, H, C and O have been applied very successfully in wide ranging situations. Improvements in the sensitivities of measurements of isotopic concentrations in small samples have increased the scope of a number of radioactive isotope dating methods. Consequently, *processes* and *time scales* in most geologic settings can now be characterized fairly well.

Soils have, however, resisted most attempts to study their dynamics, even using the most widely applied methods, and have continually challenged the imagination of physicists and chemists. The reason why soils are difficult to characterize is that they are principally composed of a host of labile compounds, such as a great variety of organic compounds, clays, oxy-hydroxides and precipitates, which all act as *open systems* in the chemical sense, freely exchanging elements, and their isotopes with the environment. The soil matrix behaves therefore as a chemically *open system*, eliminating the use of conventional nuclear methods, which must be based on well-defined chemical behaviors of the parent and daughter nuclides. But it now appears that there is a good hope of overcoming this problem, which may truly be called the last bastion in the present context.

We must point out here that there have been several earlier attempts to study soil erosion, e.g. by mass-balance methods, use of U-Th series nuclides, and soil dating with different radionuclides including the cosmogenic ^{10}Be produced by cosmic rays in the earth's atmosphere. The mass-balance method mainly provides information on the (*in situ*) weathering rates (Kirkby and Morgan, 1980; Velbel, 1985; Lerman, 1994). Most of the nuclear techniques are relevant but at best can provide only approximate answers to soil erosion and formation histories, since they must make the assumption that the soil column is a closed system with strong adsorption and retention of ^{10}Be brought into the soil by rainfall. Furthermore, these techniques, as they have been used so far, do not provide incremental soil erosion rates as a function of time, as is often needed. The cosmogenic nuclide ^{10}Be, with its fairly well known source function and long half-life (1.5×10^6 y), has afforded the most meaningful estimates to date, of soil dynamics parameters (Monaghan et al., 1983; Pavich et al., 1985, 1986; Brown, 1987a,b). In this latter method, the soil column is treated as a closed system, receiving ^{10}Be at the soil surface from rainfall, which scavenges atmospheric cosmogenic ^{10}Be. The soil erosion rate has been estimated by comparing the reduction in ^{10}Be content from that expected from its fallout (Pavich et al., 1984; Brown, 1987a,b).

The environmental tracers ^{137}Cs and $^{239-240}$Pu, produced in nuclear weapons tests and disseminated globally during the 1950s and early 1960s have proven very useful for determining soil erosion on decadal time scales (Ritchie and McHenry, 1975; Loughran et al., 1988; Quine and Walling, 1991; Quine et al., 1992). Since these transient tracers often result in characteristic depth profiles with a transition at sub-surface depths, they offer some advantages over naturally produced tracers. The latter alone, however, are potentially capable of providing long-term averaged soil erosion rates.

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The new possibility of introducing numeric time control in soil dynamics is based on the production of stable and long-lived nuclides in nuclear interactions of cosmic radiation in (i) the atmosphere, and (ii) in terrestrial solids, and their convenient detection in diverse terrestrial samples using improved conventional methods and the newly developed accelerator mass spectrometry method. The convenience and the precision with which a few million atoms of several nuclides can now be measured using accelerator mass spectrometry (AMS), has tremendously widened the scope of the field of cosmic ray produced (cosmogenic) isotopic changes in planetary studies (Elmore and Philipps, 1987; Lal, 1988, 1998) in general, and has opened up the field of soil dynamics using nuclear methods (Monaghan et al., 1983; Pavich et al., 1985, 1986; Lal et al., 1991, 1996; Barg, 1992; Barg et al., 1997).

The central goal of this paper is to establish that it is possible to quantify soil processes by using some of the cosmogenic radionuclides (Lal and Peters, 1967; Lal, 1988) as tracers. These nuclides are brought into the soil by rainfall, and also produced directly, *in situ* in soil minerals. We will confine our attention in this paper to three radionuclides, ^{32}Si (half-life ~150 y), ^{14}C (half-life = 5730 y) and ^{10}Be (half-life = 1.5x 10^6 y). The nuclides ^{10}Be and ^{14}C are conveniently measured in surface deposits using AMS, while ^{32}Si activities can be conveniently measured by milking its daughter nuclide, ^{32}P, and counting its activity using a low-level scintillation system (Morgenstern et al., 1996; Morgenstern, priv. comm.). The atmospheric nuclides ^{32}Si and ^{10}Be are removed from the atmosphere after their production and introduced in surface soils. The nuclides ^{14}C and ^{10}Be are produced in the atmosphere, but they are also produced directly, *in situ*, in the soil matrix (Fig. 1; Table 1).

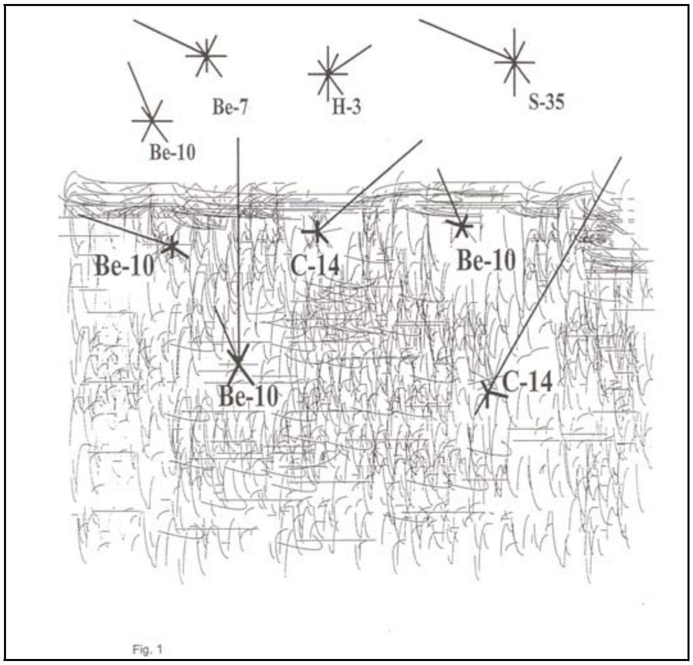

Figure 1. A large number of stable and radioactive nuclides are produced in cosmic ray interactions with the atmospheric and surficial nuclei (Lal and Peters, 1967; Lal, 1988). ^{10}Be and ^{32}Si nuclei produced in the atmosphere are introduced in soils by rainfall. Atmospheric ^{14}C finds its way to the soils in several ways. The radionuclides ^{10}Be and ^{14}C are also directly produced, *in situ*, in quartz and in other mineral grains in soils.

We briefly discuss our present knowledge of the production rates of the radionuclides listed in Table 1, as well as studies of their distribution in the atmosphere, hydrosphere and lithosphere, with special reference to soils. Finally, we present models of soil erosion and soil formation rates based on these radiotracer studies, and point out their potential for studying soil dynamics.

Production rates and distribution of cosmogenic ^{10}Be, ^{14}C and ^{32}Si in the hydrosphere and in soils

It is now feasible to fairly easily measure both the cosmogenic nuclides produced in the atmosphere, and those produced *in situ* in terrestrial solids. It has now become conventional to refer to the former class of nuclides, which can be removed from the atmosphere by aerosol scavenging by rainfall, as the *garden-variety nuclides*, in order to distinguish them from those produced directly, *in situ* in terrestrial solids by cosmic ray interactions (Fig. 1). ^{10}Be and ^{32}Si produced in the atmosphere belong to this class of *garden-variety nuclides*; ^{10}Be is also produced *in situ* in the soil matrix.

Table 1. Radionuclides of potential use in studies of soil dynamics.

Nuclide	Half-life	Principal production mechanism	
^{32}Si	~150 y	Atmosphere:	Ar spallation†
^{14}C	5730 y	Atmosphere:	neutron capture in N
		Soils:	O spallation†
^{10}Be	1.5 x 10^6 y	Atmosphere:	O, N spallation†
		Soils:	O spallation†

†By spallation, we mean fragmentation of target nuclei in nuclear interactions of energetic particles of cosmic radiation.

Atmospheric origin ^{10}Be and ^{32}Si

The cosmic ray production rates of ^{10}Be and ^{32}Si in the atmosphere have been experimentally determined. The measured global fallout rates of ^{10}Be, 4.9 x 10^{-2} atoms cm^{-2} sec^{-1} (Somayajulu et al. 1984) and 3.9 x 10^{-2} atoms cm^{-2} sec^{-1} (Monaghan et al., 1985/86) compare well with its estimated global production rate of 4.5 x 10^{-2} atoms cm^{-2} sec^{-1} (Lal and Peters, 1967). The stratospheric ^{10}Be is removed to the troposphere by large-scale air circulation within a period of ~1 y. The depositional flux of ^{10}Be is higher at the mid-latitude tropospheric breaks at 30°-40° N/S, in agreement with predictions (Lal and Peters, 1967).

The atmospheric fallout rate of ^{32}Si has also been measured extensively in wet precipitations, and in ocean waters (Kharkar et al., 1966; Lal et al., 1979; Somayajulu et al., 1991). The result based on atmospheric fallout studies, 2.5 x 10^{-6} Bq cm^{-2} (which corresponds to a flux of 5.4 x 10^{-4} atoms ^{32}Si cm^{-2} sec^{-1}, taking a value of 150 y for the half life of ^{32}Si) is consistent with its measured oceanic inventory of

4.3 x 10^{-4} Bq cm^{-2} (Somayajulu et al., 1991).

The half-life of ^{32}Si has recently been re-determined (since 1986) by several authors (see Somayajulu et al., 1991 for references, and Nijampurkar et al., 1998). These values (excluding the values of Elmore et al.. and Kutschera et al. (referred to in Somayajulu et al., 1991) which are known to be erroneous due to appreciable mass spectrometric fractionation effects), range between 133 and 178 y. We have therefore adopted a working value of ~150 y for the half-life of ^{32}Si.

In situ produced ^{14}C and ^{10}Be

A number of *in situ* produced cosmogenic nuclides have been measured in a variety of terrestrial solids (Marti and Craig, 1987; Lal, 1988a; Nishiizumi et al., 1993; Cerling and Craig, 1994). The *in situ* production rates of ^{10}Be and ^{14}C quartz have been measured accurately by studying well-documented natural samples (^{10}Be: Nishiizumi et al., 1989; ^{14}C: Lal and Jull, 1994; Jull et al., 1994). At sea level, and at geomagnetic latitudes ≥ 60°, the production rates of ^{14}C and ^{10}Be in quartz are 20 and 6.3 atoms g^{-1} quartz y^{-1}, respectively.

Nuclide data, applications and tracer models for cosmogenic nuclides in soils

Of the two *garden-variety* cosmogenic nuclides, ^{10}Be and ^{32}Si, ^{10}Be has been studied most extensively in a variety of surface soils and soil profiles. One finds a fairly good amount of its fallout retained by the soils, in the range of 10-50% of the secular equilibrium value. The first few measurements of ^{10}Be in soils were made by Moller and Wagener (1967), but real insight into ^{10}Be adsorption by soils was gained with the first detailed ^{10}Be studies made possible using accelerator mass spectrometry (AMS), by Monaghan et al. (1983). This work was soon followed by several papers by Brown and Pavich (Brown et al., 1981; Brown, 1987a,b, 1988; Pavich et al., 1984, 1985, 1986), which include studies of both surface soils and soil profiles.

In the case of ^{32}Si, its activity has been measured in rivers, ground waters, surface soils and in oceans (Nijampurkar et al., 1966; Lal et al., 1970, 1976; Somayajulu et al., 1991). The only measurements to date of ^{32}Si in soil profiles are by Frohlich (Frohlich et al., 1987). Considerable work needs to be done on ^{32}Si distribution in soils.

We will discuss first the available data and results on the garden-variety nuclei, ^{10}Be, and the *in situ* produced ^{14}C and ^{10}Be nuclei, and their potential applications with reference to particular soil models.

Garden-variety ^{10}Be in surface soils: ^{10}Be content and soil erosion rates

The application of atmospheric ^{10}Be to soil studies is based on the fact that beryllium is particle active, i.e. it is adsorbed quickly on surfaces. The distribution coefficient, K_d, for partition of dissolved Be between soil and water has been studied by You et al. (1989) in river waters, and in different soil minerals as a function of pH in the laboratory. At pH > 5, K_d values lie in the range of 3 x 10^4 – 10^6. Organic matter is found to retain ~100 times more ^{10}Be/g sediment (see Barg et al., 1997 for detailed study and discussion on the adsorption of ^{10}Be in different soil phases).

In agreement with expectations, the near-surface ^{10}Be concentrations in bulk soils typically lie in the range of 10^7-10^9 ^{10}Be atoms g^{-1}. In the Chinese loess plateau, the corresponding values for loess and paleosols at Weinan and Baoji are 2-3 x 10^8 and 4-6 x 10^8 atoms g^{-1} (Gu et al., 1996). The ^{10}Be concentration in the deposited dust is estimated to be ~ 2 x 10^8 atoms ^{10}Be g^{-1}, indicating that this is the value of surface soils in the dust source regions (Gu et al., 1996). This is close to the median value of U.S. soils sampled by Brown et al. (1981, 1988); see also Brown (1987a).

Studies of paleosols in the Chinese loess plateau (Gu et al., 1996, 1997) have revealed valuable information about soil weathering processes, as well as on incorporation of ^{10}Be in clays. In modern as well as 2.5 x 10^6 y old paleosols, the garden-variety ^{10}Be was found to adhere strongly to the clays in the B-horizon, with an estimated loss of only 5±5% ^{10}Be to the loess unit below the paleosol (Gu et al., 1996).

The history of pedogenesis in continental soil profiles, however, differs from that in the loess plateau, where dust and water deposition progresses with varying intensities of dust and water fluxes. Loess layers signify dust deposition with appreciably decreased water fluxes. In the case of continental soils, a soil profile continuously receives new precipitation, accompanied by some dust influx. (The soil profile evolves as a result of balance between erosion of topsoil, dust influx and formation of soil in the C-horizon and at the soil-bedrock interface as a result of weathering by acidic (organic) solutions originating in the O-A layers due to decomposition of organic plant matter).

The depth distribution of garden-variety ^{10}Be in soil profiles shows a great variance (Pavich et al., 1986; Pavich and Vidic, 1993). However, the main characteristics are high concentrations in the near-surface region, and decreasing concentrations at depth. There is sometimes a sub-surface maximum in ^{10}Be concentrations, which presumably refers to eluviation of (^{10}Be-rich) clays.

The content of ^{10}Be in soil profiles have generally been used to put constraints on soil ages. In the case of a China Hat soil profile, Pavich et al. (1985) have estimated the soil erosion rate. Brown et al. (1985) and Brown (1987a,b) have placed greater emphasis on the ^{10}Be data being more relevant to soil erosion rates than soil ages. Brown (1987a) observed a generally valid inverse relationship between ^{10}Be content in the soil and erosion rate. Barg et al. (1997) showed that the ^{10}Be soil content is primarily dictated by soil erosion, in conformity with the studies of Brown (1987a). The cumulative ^{10}Be inventories in a soil profile must clearly be related to their ages, but not necessarily in a unique fashion, because on the one hand, the penetration of ^{10}Be to depths depends on the soil porosity, texture and climate, the factors which determine how efficiently the acidic solutions in the O-A layers permeate downward (which determine the efficiency of weathering of the bedrock), and on the other hand to the rate of soil erosion from the surface.

To quantify these concepts, a material balance model for incorporation of atmospheric ^{10}Be in soils is presented below. A plausible two-layer model for the balance of soil soil and garden-variety radiotracers carried by rainfall on to the soil surface is considered (Fig. 2).

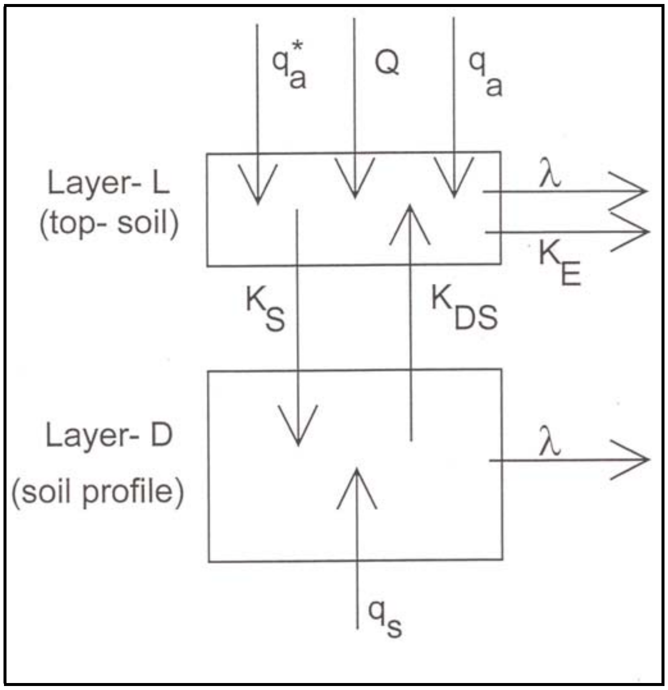

Figure 2. An operational model for the balance of soil and garden-variety ^{10}Be in a soil profile. The first order rate constants K, and other parameters are defined in the text (Sec. 3a).

- S layer: the topsoil layer comprising O and A horizons.
- D layer: remainder of the soil profile comprising B and C-horizons.
- z_0: thickness of S layer (cm).
- z: thickness of D layer (cm).
- ρ: average soil density (g cm^{-3}).
- q_a: aeolian flux of dust (g cm^{-2} y^{-1}).
- Q: flux of atmospheric ^{10}Be in topsoil (atoms cm^{-2} y^{-1}).
- q^*_a: flux of accreted aeolian ^{10}Be (atoms cm^{-2} y^{-1}).
- q_s: soil formation by weathering rate in the D layer (g cm^{-2} y^{-1}).
- N^*_S: ^{10}Be inventory in the S layer (atoms cm^{-2}).
- N^*_D: ^{10}Be inventory in the D layer (atoms cm^{-2}).
- K_S: first order rate constant for transfer of tracer from S to D layer.
- K_{DS}: first order rate constant for transfer of soil from D to S layer.
- K_E: first order rate constant for removal of soil from S layer.
- λ: nuclear disintegration constant for the radiotracer.

In steady-state, following relations can be developed from material balance considerations for garden-variety ^{10}Be:

$$q_s = z\,\rho\,K_{DS} \tag{1}$$

$$q_s + q_a = z_0\,\rho\,K_E \tag{2}$$

$$N^*_S\,(\lambda+K_E+K_S) = Q + q^*_a + N^*_D\,K_{DS} \tag{3}$$

$$N^*_S K_S = N^*_D\,(\lambda+K_{DS}) \tag{4}$$

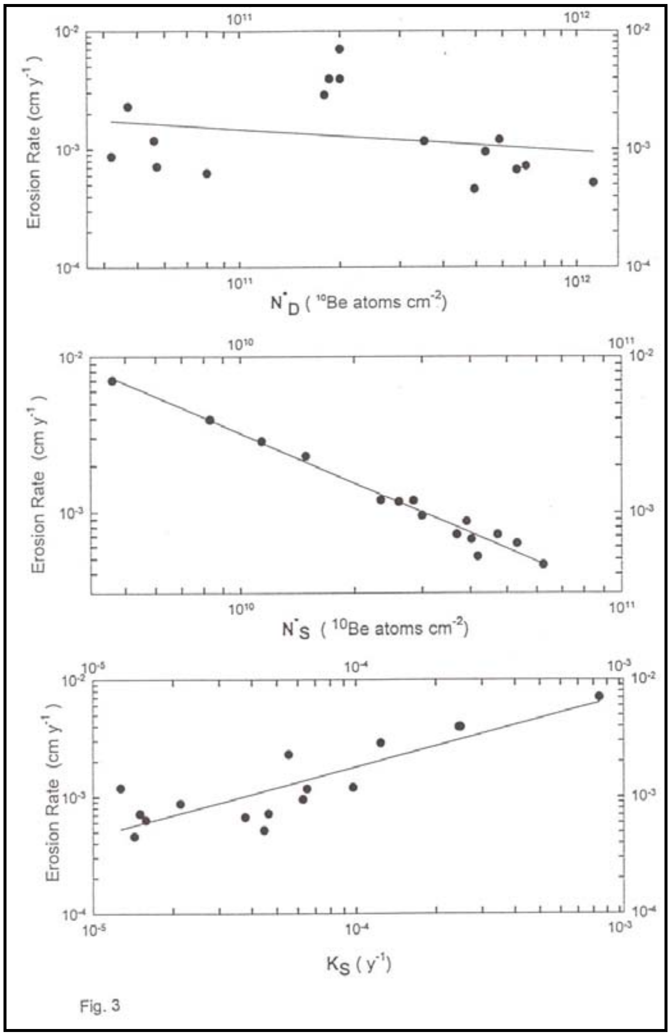

Figure 3. Model soil erosion rates and ^{10}Be soil penetration rates, based on ^{10}Be data for soils from Merced River and Sava River terraces, using equations 5 and 6. ^{10}Be data are from Pavich et al. (1986) and Pavich and Vidic (1993).

The principal unknowns in these equations are K_E and K_S, which can be evaluated from measurements of N^*_S and N^*_D. The rate of ^{10}Be addition to the soil, Q, is fairly well known, and fairly good estimates of q_a can be made from geochemical data. The following relations can be developed from equations (1-4), for K_E and K_S in terms of N^*_S and N^*_D:

$$K_E = \frac{N_D^*}{N_S^*}\left[\frac{Q+q_a^*}{N_D^*} - \lambda\right] - \lambda \tag{5}$$

$$K_S = \frac{N_D^*}{N_S^*}\left[\lambda + \frac{z_0}{z}K_E\right] \tag{6}$$

with the corresponding erosion rate, E (cm y^{-1}), being equal to z_0 K_E.

To treat the limited available data on ^{10}Be in soils, we will neglect the aeolian influx of dust and ^{10}Be, which may however be appreciable in some cases. The present annual depositional fluxes of dust in Miami (Florida, USA) have been estimated to be about 10^{-4} g cm^{-2} y^{-1}, which agrees well with the estimates for aeolian dust deposition in the tropical North Atlantic (Prospero, 1999). The ^{10}Be concentrations of clay-rich dust in Weinan and Baoji have been determined to be ~2 x 10^{8} atoms g^{-1} (Gu et al., 1996) which is within the range of surface soils in USA (Brown et al., 1987a,b). This leads to an average ^{10}Be depositional flux of 2 x 10^{4} atoms cm^{-2} y^{-1} via aeolian dust, about two orders of magnitude smaller than the global depositional flux of 1.4 x 10^{6} atoms ^{10}Be cm^{-2} y^{-1}. The soil erosion rates are estimated to generally lie between 2-5 tons ha^{-1}, i.e. (2-5) x 10^{-2} g cm^{-2} y^{-1}. The aeolian mineral dust deposition rate cited earlier is two orders of magnitude smaller; however, we note that aeolian flux of dust may be significant in some settings.

Sufficient ^{10}Be data are available for its concentrations in soil profiles to obtain a fair picture of its characteristics in soils, with special reference to its recharge rate in the profile, and typical variations in soil erosion rates. We first consider the data for ^{10}Be distributions in soils from terraces: (i) 5 profiles from Merced River, California (Pavich et al., 1986), and (ii) 11 Alpine glacio-fluvial profiles from Sava River, Slovenia (Pavich and Vidic, 1993). For our calculations we adopted the following values for the parameters: average soil density, r = 1.7 g cm^{-3}; Q = 1.4 x 10^{6} atoms ^{10}Be cm^{-2} y^{-1}, z_0 = 25 cm. The following conclusions emerge from these 16 profiles:

I. The degree of incorporation of ^{10}Be in these soil profiles, compared to its secular equilibrium value, lies between 1.5-40%.

II. The estimated soil erosion rates using equation (5) lie between 5 x 10^{-4} and 7 x 10^{-3} cm y^{-1} (Fig. 3).

III. There exists a very weak negative correlation (Fig. 3) between the estimated erosion rates, E and the ^{10}Be inventory in the soil profile, N^*_D; E = const. $(N^*_D)^{-0.19}$. N^*_D and N^*_S are also very poorly related in these soils.

IV. The values of K_S, which represents the rate of recharge of ^{10}Be to the deeper soil layers in the profile, lie between 1.3 x 10^{-5} and 8.4 x 10^{-4} y^{-1} (Fig. 3).

V. The parameters K_S, N^*_S and E are closely related (see Fig. 3)

We will now consider the erosion rates for the six soil profiles which were studied by Barg et al. (1997), analyzing both ^{10}Be and ^{9}Be concentrations in bulk soil as well as in soil minerals. Consequently in these soil profiles, the soil formation rates are also known, based on the depth variation of ^{10}Be/^{9}Be ratios in the soil minerals (see section 3b). The estimated soil erosion rates and soil formation rates are listed in Table 2, which show the same trends as the data for erosion rates and K_S in Fig. 3.

The soil erosion rates are seen to be greater than the soil formation rates by factors of 2 to 100. The high values are apparently a direct consequence of the much lower values of N^*_S in these profiles, which could have resulted from an imbalance between natural soil formation rate and enhanced soil erosion rates in recent periods.

^{10}Be/^{9}Be ratio chronometer for in situ weathered soil profiles

Considering the mechanics of soil erosion, Lal et al. (1991) proposed a chronometric method for determining soil formation rates based on the garden-variety ^{10}Be. The weathering of the soil at the bedrock-soil interface, and in the C-horizon, occurs due to percolation of acidic organic solutions in the O and A layers, which happens primarily in wet periods after dry spells, by rapid percolation of fluids through cracks in the argyllic horizon in the B layer. Consequently the ^{10}Be present in the O and A layers gets fairly well homogenized with ^{9}Be, over periods of the order of few thousand years in the top-soil, and the ratio ^{10}Be/^{9}Be is also fairly well fixed within narrow limits. Beryllium is fixed in the clay minerals formed during weathering in the C-horizon. As the clay content increases, their ^{10}Be decays with time, and therefore the ^{10}Be/^{9}Be ratio in clays decreases upwards, depending on the rate of formation of clays. In this model, the highest ^{10}Be/^{9}Be ratios in clays will be found at the bottom of the soil profile, with ratios decreasing towards the surface, as one goes into and within the B-horizon. It then follows that the ^{10}Be/^{9}Be ratios in clays should serve as a chronometer for the ages of soils. This model is depicted in a cartoon of a soil profile in Fig. 4.

Table 2. Results for model soil erosion and soil formation rates for six soil profiles studied for ^{10}Be and ^{9}Be concentrations in bulk soils and in different minerals, by Barg et al. (1997).

		^{10}Be activity					
Sample Code	**Core Length (cm)**	**in the top-soil (atoms g^{-1})**	**in the soil column (atoms cm^{-2})**	**K_E**	**Erosion rate (cm y^{-1})**	**K_S**	**Soil formation rate (cm y^{-1})**
Great Valley (GV)	200	5.0×10^{8}	2.7×10^{11}	6.0×10^{-5}	1.5×10^{-3}	1.0×10^{-4}	4.1×10^{-5}
Slovenia M6	135	3.6×10^{8}	4.7×10^{10}	9.0×10^{-5}	2.2×10^{-3}	5.2×10^{-5}	2.5×10^{-5}
Slovenia M4	309	1.0×10^{9}	5.3×10^{11}	2.7×10^{-5}	6.7×10^{-4}	3.3×10^{-5}	1.3×10^{-4}
China Hat (CH)	350	5.0×10^{8}	5.3×10^{11}	5.4×10^{-5}	1.4×10^{-3}	1.1×10^{-4}	1.4×10^{-4}
South Mali Ferricrite (SM)	1200	1.0×10^{7}	2.5×10^{11}	3.0×10^{-3}	7.6×10^{-2}	3.7×10^{-2}	3.2×10^{-4}
Ivory Bauxite (IC)	1000	1.0×10^{7}	1.2×10^{11}	3.2×10^{-3}	7.9×10^{-2}	2.2×10^{-2}	2.2×10^{-4}

Reference is made to Barg et al. (1997) for details of the soil profiles.

To test the validity of the model, Lal et al. (1991) examined more than 60 surface soils for their ^{10}Be/^{9}Be ratios. Subsequently Barg (1992) and Barg et al. (1997) studied six global soil profiles from different climatic regimes for their ^{10}Be/^{9}Be ratios. These soils include slightly alkaline, acidic leached, and oxide-rich varieties. The results validated the expectations for the ^{10}Be/^{9}Be chronometer model (Fig. 5), and yielded soil ages in the range of 50,000 years to 10 million years for the six selected soil profiles (Barg et al., 1997).

Soil erosion rates based on in situ produced ^{14}C and ^{10}Be in quartz grains

In the case of *in situ* weathered soil profiles, erosion rates can generally be studied by using the *in situ* cosmic ray produced radionuclides ^{10}Be and ^{14}C. The basis of this model is their continued production by cosmic ray interactions in quartz grains during weathering. (Since the bedrock is at a considerable depth from the surface (a meter or more in typical cases), the quartz grains are initially well shielded from cosmic radiation. However, as the quartz grains move upwards with the march of evolution of the soil profile, they acquire considerable amounts of *in situ* ^{14}C by cosmic ray irradiation). The depth profile of ^{14}C in quartz grains depends only on the rate of erosion of soil profile.

As a unique variant of this model, we studied a soil profile in Reston, Virginia, which had a quartz dyke cutting into it. In this profile, nature provided a profuse supply of quartz in the dyke, which weathered and fragmented as the soil eroded. The estimated erosion rates based on measurements of *in situ* cosmogenic ^{14}C and ^{10}Be are shown in Table 3.

As can be seen from Table 3, the results based on the radionuclide ^{14}C are not consistent with those based on ^{10}Be. The ^{14}C data show a high erosion rate, $\geq 3 \times 10^{-3}$ cm y-1 in the past $\leq 5.7 \times 10^{3}$ y, but the ^{10}Be on the other hand yield an erosion rate of $\leq 3 \times 10^{-4}$ cm y-1 in the past $\geq 1.8 \times 10^{5}$ y. These data are, however, quite consistent with a model wherein the quartz vein first eroded at the slower of

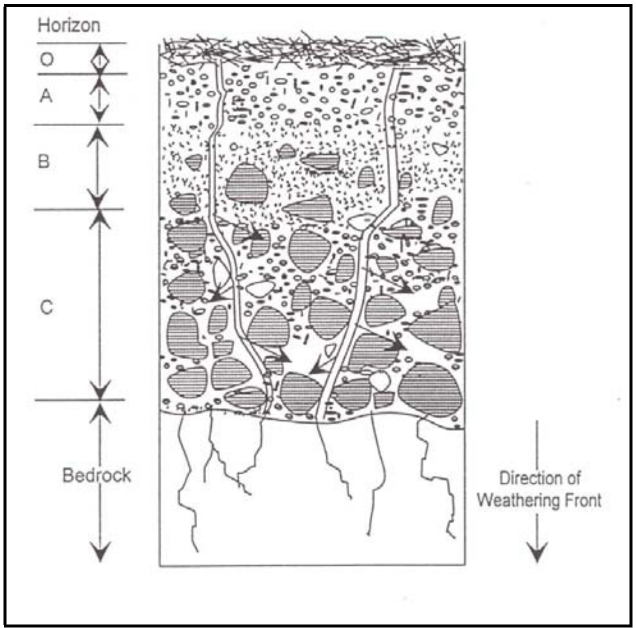

Figure 4. A schematic depicting the essentials of the ^{10}Be/^{9}Be chronometer model (Lal et al, 1991). The garden-variety ^{10}Be is equilibrated initially in the organic matter-rich O and A layers. The evolution of the soil profile occurs when the acidic solutions in O and A layers percolate through channels in the soil profile, weathering the matrix in the C horizon and the bedrock surface. As the weathering front moves downward, ^{10}Be concentrations in clays progressively decrease due to radioactive decay, as they move upwards to and through the B-horizon (figure adapted from Barg et al., 1997).

Table 3. Inferred *in situ* ^{14}C and ^{10}Be model erosion rates and effective exposure ages of the Reston (Virginia, USA) soil profile (Lal et al., 1996).

Nuclide	Mean erosion rate (cm y^{-1})	Effective exposure age (y)
^{14}C	$\geq 3 \times 10^{-3}$	$\leq 5.7 \times 10^{3}$
^{10}Be	$< 3 \times 10^{-4}$	$\geq 1.8 \times 10^{5}$

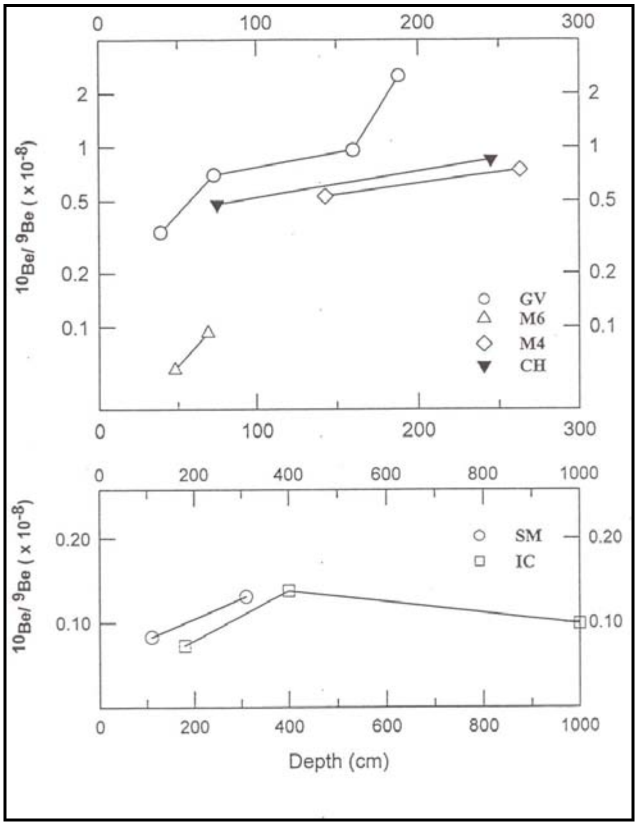

Figure 5. Depth variation of ^{10}Be/^{9}Be ratios in authigenic mineral phases in six soil profiles: Great Valley (GV), two Slovenia soils (M6 and M4), China Hat (CH), South Mali ferricrite (SM), and Ivory Coast bauxite (IC). The upper diagram is for the ^{10}Be/^{9}Be ratios in the clay fractions, the lower diagram for the ferric hydroxide phase in oxide-rich soils (after Barg et al., 1997).

the two erosion rates, and was then buried abruptly or slowly with a thickness of soil exceeding 1.3 m, to result in the preferential decay of accumulated ^{14}C over ^{10}Be in the vein. The soil burial period has to be greater than two mean lives of ^{14}C, > 2 x 10^4 y to lower the $^{14}C/^{10}Be$ ratio significantly.

However, upon going through the geological history of the profile, the erosional model suggested by the radionuclides is found to be quite acceptable, as is discussed in detail by Lal et al. (1996).

Erosion and formation rates of cultivated soils

So far, we considered soil formation and erosion processes only for in-situ weathered soil profiles. It is quite manifest that the cosmogenic nuclides can also serve as tracers for constraining soil dynamics for tilled or re-deposited soils, since their source functions are known, and one can also make use of the information obtained from studies of naturally weathered soils to constrain distribution of cosmogenic tracers prior to cultivation/re-deposition. To illustrate this potential of cosmogenic tracers, we briefly consider how they can be used for determining erosion rates of cultivated soils, with special reference to the soils in the corn-belt in USA. In this case, a fairly good idea exists about the geologic history of the cultivated soils prior to their cultivation. From a preliminary study of the lithology and the history of soil in the corn-belt, it appears that the garden-variety ^{10}Be is best suited to determine erosional losses of solubles and solids from the site. From material balance considerations (such as in Sec. 3a), one can also hopefully directly constrain soil erosion rates. One has to also consider soil gains due to the new soil produced due to weathering of the loess deposited during the past ~20 ky. This can be estimated by examining ^{10}Be concentrations and $^{10}Be/^{9}Be$ ratios in the soil matrix at the loess-soil interface. From surveys of garden-variety ^{10}Be concentrations, and $^{10}Be/^{9}Be$ ratios in clay minerals in the cultivated zone and in the layers below the cultivated zone, one can put constrains on the present day soil formation rates. If the estimated erosion rates exceed the soil formation rates, a safe limit to the duration of viable cultivation can be projected! If opposite is the case, cultivation at the present rate could be sustained!

CONCLUSIONS

We have discussed some special applications which the cosmogenic nuclides, the garden-variety (atmospheric) ^{10}Be and the *in situ* produced ^{10}Be and ^{14}C have found in studies of soil ages, soil formation, and erosion rates. This field is in its infancy as far as tracer applications to soil dynamics is concerned. However, sufficient groundwork has been done in determining the cosmogenic tracer production mechanisms, and their production rates. The garden-variety ^{10}Be has been studied fairly extensively in a wide variety of soils with a view to determine its concentrations in surface soils, as well as its content in several soil profiles. Six soil profiles from widely different climatic regimes were studied with a view to explore the feasibility of using the $^{10}Be/^{9}Be$ ratio in authigenic soil minerals as a "chronometer" for determining soil ages (Lal et al., 1991; Barg et al., 1997). This method shows considerable promise since the initial $^{10}Be/^{9}Be$ ratio is fixed within rather narrow limits in the O-A layer. The acidic solutions with this $^{10}Be/^{9}Be$ ratio introduce this ratio in the clay minerals formed in the C-horizon, and at the soil-bedrock interface.

We have also presented material balance models for soils and cosmogenic radiotracers, and show how one can estimate both the rate of recharge of tracer in the soil profile, and the soil erosion rates. These estimates are presented for 16 soil profiles from river terraces and for 6 soil profiles studied for testing the $^{10}Be/^{9}Be$ ratio chronometer.

We also show that the cosmogenic *in situ* ^{10}Be and ^{14}C produced in quartz grains in *in situ* weathered soil profiles hold promise for estimation of soil erosion rates. The case study of erosion of a soil profile from Reston, Virginia with a quartz dyke using the two radionuclides ^{10}Be and ^{14}C provided information on soil erosion rates during the recent 6 ky and in the past 180 ky periods. This application is based on continued exposure of cosmic ray shielded quartz grains to greater intensity of cosmic radiation, with the progression of soil erosion. (This method is clearly not applicable for soils deposited by fluvial transport.)

Obvious future applications of the cosmogenic tracers would be for dating paleosols, for determining fairly accurate erosion rates of soils under natural conditions of evolution in different climatic regimes, and finally, towards extending these studies to cultivated soils. Research on soils has so far been on purely academic grounds, to learn about the geochemistry of atmospheric ^{10}Be, and to study the potential applications of the *in situ* nuclides. We believe that during the next few decades, we shall see much more of cosmogenic soil studies that are application-oriented.

ACKNOWLEDGMENTS

The author would like to take this opportunity to acknowledge earlier collaborations with Eunjoo Barg, Marc Caffee, Robert Finkel, A.J.T. Jull, Milan Pavich and John Southon in exploratory studies to determine the potentials of the cosmogenic radionuclides. The critical comments and valuable suggestions by Derek Fable on this manuscript are gratefully acknowledged. I am grateful to late Mr. Myron Eichen for his interest in the present studies and for generous support of the present research. Thanks are also due to Richard Morgan for helping to bring the manuscript to its present form.

Finally, we are grateful to Mark Nearing and Darrell Norton for encouraging us to present results of our soil studies at this conference.

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