^{10}BE, ^{14}C DISTRIBUTION, AND SOIL PRODUCTION RATE IN A SOIL PROFILE OF A GRASSLAND SLOPE AT HESHAN HILLY LAND, GUANGDONG

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ABSTRACT. Concentrations of organic carbon, carbon isotopes (^{13}C and ^{14}C), atmospheric ^{10}Be in soil, and in situ ^{10}Be in bedrock and weathering rock were determined in a study of a profile of a grassland slope at the Heshan Hilly Land Interdisciplinary Experimental Station, Chinese Academy of Sciences, in Guangdong Province, China. A good linear relationship between depth and the ^{14}C apparent age of the organic carbon demonstrates that the rock weathering process and the accumulation process of organic matter in the slope are relatively stable. Both ^{14}C and ^{10}Be results show that about 34% of soil in the grassland slope has been eroded during the past 3800 yr. The ^{10}Be results for interstitial soil from weathered rocks show that the 90-cm-thick weathering rock layer above the bedrock has evolved over a period of 1.36 Myr. The concentrations of in situ ^{10}Be in the weathered rock and bedrock are 10.7×10^4 atoms/g and 8.31×10^4 atoms/g, respectively. The weathering rate of the bedrock, equivalent to the soil production rate, was estimated at 8.8×10^{-4} cm/yr, and the exposure ages of the weathered rock and the bedrock were 72 kyr and 230 kyr, respectively.

INTRODUCTION

Soil contains abundant paleoclimatic information and is also an important source or sink for major greenhouse gases such as carbon dioxide, methane, and nitrogen oxides. With respect to global change, interaction between the pedosphere and other spheres of the crust has attracted more and more attention from researchers. The key to understanding soil development lies in the rate of pedogenesis and the rate of soil erosion, which requires precise dating. Only after the timescale for the development of soil is established can the paleoclimatic and paleoenvironmental information revealed by the geochemical evolution of nuclides and elements in the soil be interpreted.

Since soil is developed from the accumulation of erosional products or sedimentary deposition above the bedrock, the age of soil can be determined from the ages of the erosional surface and the sediments. The ^{14}C method has already been used in soil dating and in quantitative discussion of soil development processes (Goh et al. 1975; Trumbore 1993). ^{10}Be is a product of spallation nuclear reactions between cosmic ray particles and atmospheric gases (O, N), and has a half-life ($T_{1/2}$) of 1.5 Myr. Within a short time after its production, ^{10}Be becomes attached to aerosols and is deposited on the earth's surface via precipitation or dust fallout. ^{10}Be can be produced also through spallation reactions between cosmic ray particles and rocks on the earth's surface. The former is called "atmospheric ^{10}Be" and the latter is called "in situ ^{10}Be." In soils, Be exists mainly as oxide, or in the alkaline environment as complex anions such as $Be(OH)CO_3^-$ or $Be(CO_3)_2^{2-}$. Organic matter can also complex Be, which in part explains why Be concentration in soils is relatively high (Pavich et al. 1984). The geochemical properties and the distribution of ^{10}Be in soil have been extensively investigated in relation to the emerging field of soil dynamics and dating. Recently, researchers have applied either an open-system model (Monaghan et al. 1983; Pavich et al. 1984, 1986; McKean et al. 1993) or a closed-system model (Lal 1991; Barg et al. 1997; Heimsath et al. 1997) to estimate the

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age of soil formation using ^{10}Be. In the closed-system model, it is assumed that ^{10}Be has been locked up in authigenic minerals since soil formation, so that ^{10}Be decay can be used to determine the formation age of the authigenic minerals.

This research focuses on a soil profile of a grassland slope in Guangdong Province, China. Both atmospheric ^{10}Be in soil and in situ ^{10}Be in bedrock and weathering rock were determined, as well as organic carbon in soil and its ^{14}C and ^{13}C composition.

SAMPLING SITES AND SOIL SAMPLES

The sampling sites are located in a grassland slope in the catchment area at the Heshan Hilly Land Interdisciplinary Experimental Station, Chinese Academy of Sciences (HHLIES-CAS) (22°41′N, 112°54′E) at an elevation of about 92 m asl and slope of 18.5°. The geomorphic feature of the hilly land is convex and the underlying bedrocks are sandy shale. The land is covered by vast stretches of lawn mixed with sparsely populated shrubs. In the summer of 1999, the authors excavated 3 soil profiles along 1 side of the watershed: Profile I is located in the upper part of the slope; Profile II in the middle of the slope, about 20 m away from Profile I; and Profile III in the lower part of the slope, about 50 m further away from Profile I.

This paper deals only with investigations on Profile I. Figure 1 shows that Profile I is comprised of 3 parts: a soil layer (60 cm), a weathering layer (90 cm), and bedrock. The soil layer is itself comprised of 4 layers, while the weathering layer consists mainly of rock fragments of different sizes, with weathering clays filling the interstices. With increasing depth, this layer contains less clay and a greater amount of weathered rock fragments, the fragments becoming larger in volume. For ^{10}Be and $\delta^{13}C$ determination, the sampling interval is 2 cm above 20 cm depth, 5 cm for 20–60 cm depth, and 20 cm or 30 cm for 60–150 cm depth. One sample of the bedrock was taken. For ^{14}C determination, the sampling interval is, on average, 4 cm above 20 cm depth and 10 cm or 20 cm for 20–60 cm depth (Figure 1).

EXPERIMENT AND RESULTS

^{14}C Determination

After air-drying and picking out visible roots and fragmentary stone, all soil samples were passed through a 1-mm sieve to remove rootlets and coarse sand. All samples were treated with 0.1N HCl to remove carbonates and then rinsed with distilled water repeatedly until they became neutral. Soil samples were dried prior to analysis of ^{14}C, ^{13}C, and organic carbon. The samples for ^{14}C measurements were further ground and put into a quartz tube, where they were combusted in an oxygen stream at 800 °C to produce CO_2. The CO_2 was purified repeatedly using dry ice-acetone and liquid-nitrogen traps and then converted into Li_2C_2 in a Li-reactor under vacuum at 900 °C. Li_2C_2 was hydrolyzed into C_2H_2, which was then synthesized into C_6H_6 under catalysis. ^{14}C measurements were carried out using a 1220 Quantulus ultra low-level liquid scintillation spectrometer with a scintillation cocktail teflon vial.

^{13}C Determination

The ^{13}C content for soil samples is usually expressed as $\delta^{13}C$:

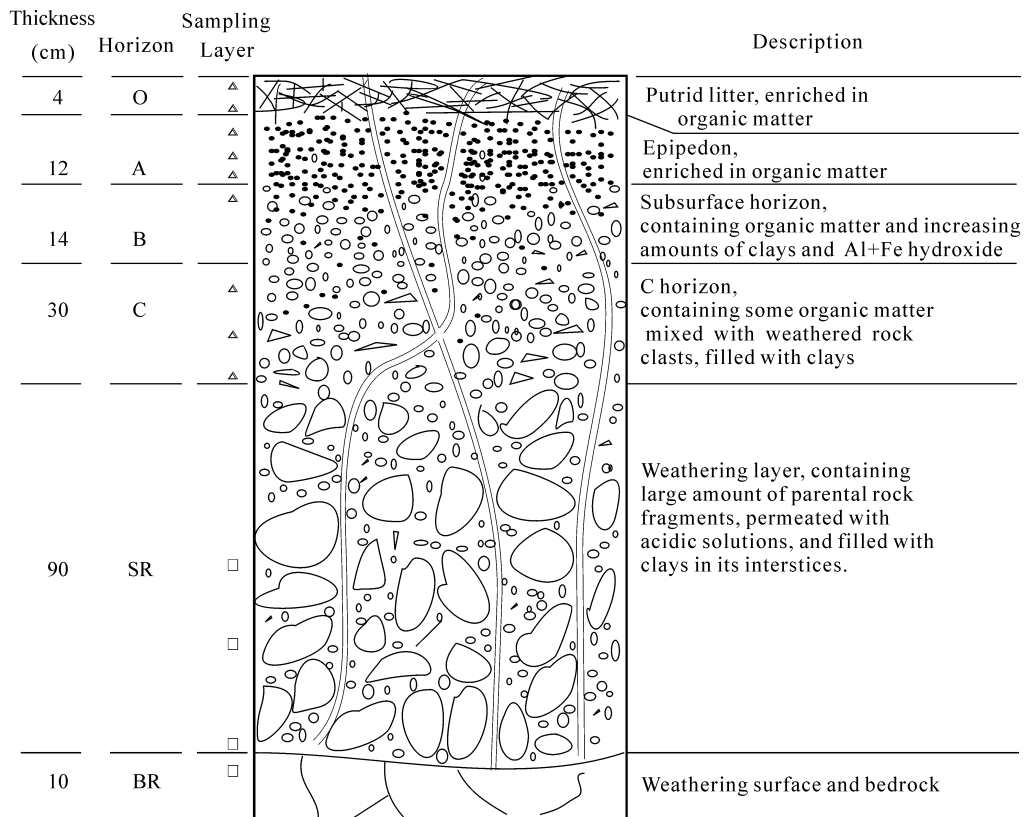$$\delta^{13}C = [(^{13}C/^{12}C)_{sample} / (^{13}C/^{12}C)_{standard} - 1] \times 1000,$$

Figure 1 Sketch of Profile I of a grassland slope at Heshan Hilly Land, Guangdong. Δ = the sampling layer for ^{14}C; ☐ = the sampling layer for in situ ^{10}Be.

where $(^{13}C/^{12}C)_{sample}$ is the isotope ratio for organic carbon in soil samples and $(^{13}C/^{12}C)_{standard}$ is the carbon isotope ratio for the PDB standard. All ^{13}C measurements were completed using the Finigen Model-251 mass spectrometric system in the Xi'an State Key Laboratory of Loess and Quaternary Geology. The modern mass spectrometry has a δ^{13}C measurement precision of ±0.02‰ and a sample preparation error of generally less than ±0.2‰. Therefore, when the difference in the δ^{13}C value for soil samples is less than ±1‰, special attention should be paid to the interpretation of the δ^{13}C data. In this work, the δ^{13}C value ranges from −27.6‰ to −15.7‰, while the average δ^{13}C values for the various horizons are −27.4‰ for Layer O, −23.3‰ for Layer A, −18.5‰ for Layer B, and −16.7‰ for Layer C.

Organic Carbon Determination

Soil samples were first oxidized into CO_2 in a vacuum system, then CO_2 volume and pressure were measured and converted into organic carbon concentration. All results are listed in Table 1 and illustrated in Figure 2. The organic carbon concentrations for the various soil horizons are as follows: 4.29% for Layer O; 1.22–0.58% for Layer A; 0.35–0.37% for Layer B, which tends to be stable; 0.21–0.16% for Layer C, which still remains stable despite the concentration being rather low.

Table 1 The concentrations of ^{10}Be, ^{14}C, ^{13}C, and organic carbon in Profile I of a grassland slope at Heshan Hilly Land, Guangdong.

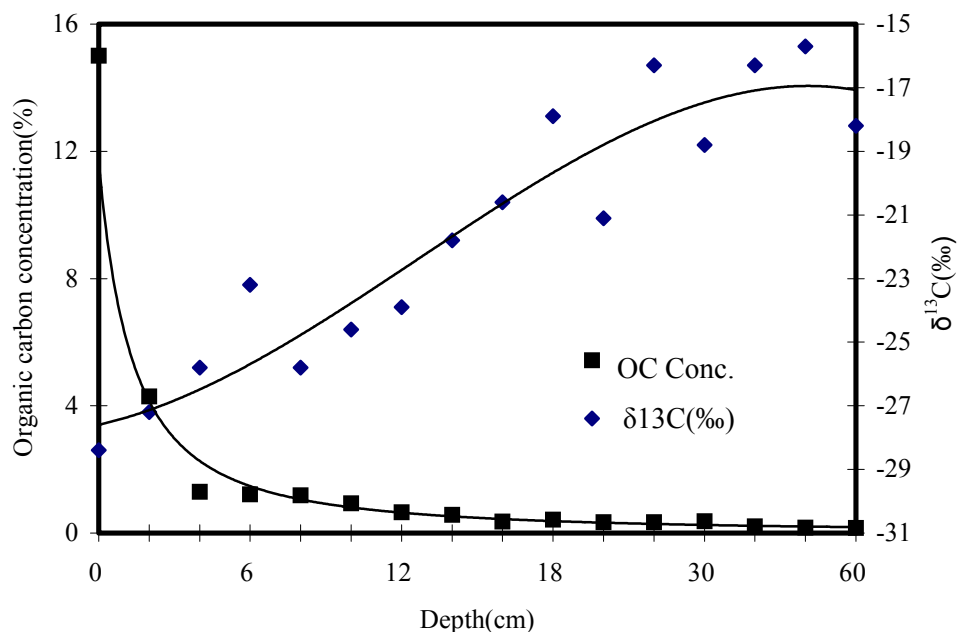| Lab code	Field code	Horizon	Depth (cm)	Organic carbon (%)	Δ^{14}C (‰)	^{14}C apparent age (BP)	δ^{13}C (‰)	Atmospheric ^{10}Be (10^8atoms/g)	In situ ^{10}Be (10^4 atoms/g)
GC-001037	HSC1-0	Litter	0	—	126	modern	−28.4	—	—
GC-001038	HSC1-1	O	2	4.29	100	modern	−27.2	0.048	—
GC-001039	HSC1-2	O	4	1.30	—	—	−25.8	0.051	—
GC-001040	HSC1-3	A	6	1.22	—	—	−23.2	0.071	—
GC-001041	HSC1-4	A	8	1.19	62	modern	−25.8	0.084	—
GC-001042	HSC1-5	A	10	0.94	—	—	−24.6	0.102	—
GC-001043	HSC1-6	A	12	0.65	—	811 ± 70	−23.9	0.112	—
GC-001044	HSC1-7	A	14	0.58	—	—	−21.8	0.094	—
GC-001045	HSC1-8	A	16	0.36	—	1032 ± 75	−20.6	0.166	—
GC-001046	HSC1-9	B	18	0.43	—	—	−17.9	0.141	—
GC-001047	HSC1-10	B	20	0.35	—	1770 ± 75	−21.1	0.13	—
	HSC1-11	B	25	0.35	—	—	−16.3	0.164	—
GC-001048	HSC1-12	B	30	0.37	—	—	−18.8	0.152	—
	HSC1-13	C	35	—	—	—	—	0.162	—
GC-001049	HSC1-14	C	40	0.21	—	2210 ± 90	−16.3	—	—
	HSC1-15	C	45	—	—	—	—	0.161	—
GC-001050	HSC1-16	C	50	0.17	—	3080 ± 105	−15.7	0.16	—
	HSC1-17	C	55	—	—	—	—	0.159	—
GC-001051	HSC1-18	C	60	0.16	—	3840 ± 110	−18.2	0.244	—
	HSC1-19	SR	80	—	—	—	—	0.119	—
	HSC1-20	SR	100	—	—	—	—	0.125	10.7
	HSC1-21	SR	120	—	—	—	—	0.073	6.4
	HSC1-22	SR	150	—	—	—	—	0.097	8.3
	HSC1-23	BR	160	—	—	—	—	—	8.13

Figure 2 Relationships of organic carbon concentration versus depth and stable carbon isotope versus depth for Profile I.

[10]Be Determination

Atmospheric [10]Be

After removal of roots and large detrital stones, the soil was desiccated under natural ventilation conditions for 24 hr and screened with a 710-μm sieve. The screened soil samples were ground in an agate mortar to form a homogeneous mixture. One gram of sample was weighed and placed in a 100-mL plastic beaker, to which a 1-mL [9]Be carrier solution (containing 0.254 mg of [9]Be) was added. Organic matter was removed with H_2O_2, then 6M HCl was added for leaching (Shen et al. 1992). The leaching solution was placed in a teflon beaker and evaporated to dryness on an electric hot plate. The residue was dissolved in 1M HCl and then transferred to an ion exchange column (DOWEX-50W×8, 100- to 200-mesh size) which was washed 10 times with 10 mL 1M HCl. The [10]Be-bearing residue was preserved in the effluent, then heated to reduce the volume to 20 mL, NH_4OH solution was added, and $Be(OH)_2$ was separated out. In a 900 °C quartz oven coated with gold, $Be(OH)_2$ was oxidized to BeO, which was then mixed homogeneously with copper powder and pressed into [10]Be targets,—just as for accelerator mass spectrometry (AMS) measurements— using a 5T target-preparation machine. All [10]Be measurements were carried out in the ETH/PSI AMS system in Switzerland. The blank [10]Be/[9]Be ratio is 10^{-14}, which is about 2 orders of magnitude lower for the soil sample. At least 2 measurements were made for each sample and the measurement precision (1 σ) was 3–5% (Suter et al. 1984).

In situ [10]Be

About 30 g of quartz grains were carefully separated from the rock samples and placed in a 500-mL plastic bottle, to which was added 300 mL of 5% HF solution. The plastic bottle was placed and fixed on a vibrator and vibrated at a rate of 200 times per min for 24 hr. The supernatant liquid was discarded and the remaining quartz grains rinsed with deionized water 5 times. Then, 300 mL of 5% HF solution was added and the quartz grains were vibrated and rinsed for the second time. The well-rinsed quartz grains were dried and 20 g weighed for analysis. A 5-ml 40% HF solution and 1-ml [9]Be carrier (containing 1.0 mg of [9]Be) were added to a teflon bottle containing 20 g of quartz; the bottle was placed on a hot plate, and the solution evaporated at 80 °C. Even though the quartz grains were completely altered, residues of other minerals might remain. After adding 2 mL of 1M HCl into the dried sample, the component containing in situ [10]Be was completely dissolved, and the residues and the solution were thus separated centrifugally. The separated solution was loaded on to the DOWEX-50W×8 (100- to 200-mesh) cation exchange resin column and washed 10 times with 10 mL of 1M HCl. A 100-mL effluent was collected in a teflon bottle. The effluent was evaporated to 10 mL on a hot plate and its pH value was adjusted to 8–9 with NH_4OH to precipitate $Be(OH)_2$. It was then oxidized at 900 °C to obtain BeO, which was mixed with an appropriate amount of copper powder and pressed into [10]Be targets for AMS measurement at the ETH/PSI AMS system in Switzerland. S555 was used as [10]Be standards in the in situ [10]Be determinations with the blank [10]Be/[9]Be ratio of 10^{-14}. At least 3 measurements were made for each sample and standard sample and the measurement precision (1 σ) was 5% (Kubik et al. 1998; Ivy-Ochs et al. 1995). The measurement results are listed in Table 1.

DISCUSSION

[14]C Apparent Age

The 0–4-cm section of Profile I is comprised mainly of litter and humic substances with ^{14}C radioactivity consistently greater than the standard value for modern carbon, demonstrating that it has

already been affected by atmospheric nuclear tests. Comparisons with the radioactivity of atmospheric ^{14}C show that the organic carbon originated from plants accumulated in the past 10 yr. As illustrated in Table 1, the penetration depth of bomb ^{14}C is 8–10 cm. Layer B (16–30 cm) has a pedogenesis duration estimated to be at least 700 yr based on the ^{14}C results. Layer C (30–60 cm) has an apparent age for the organic carbon at its base of 3840 ± 110 BP. Figure 3 illustrates the relationship between the ^{14}C apparent age for Profile I and its burial depth. One can see that there is a good linear relationship between the depth and the ^{14}C age for organic matter, which reveals that the weathering of rocks in the slope and the accumulation of organic matter are both stable.
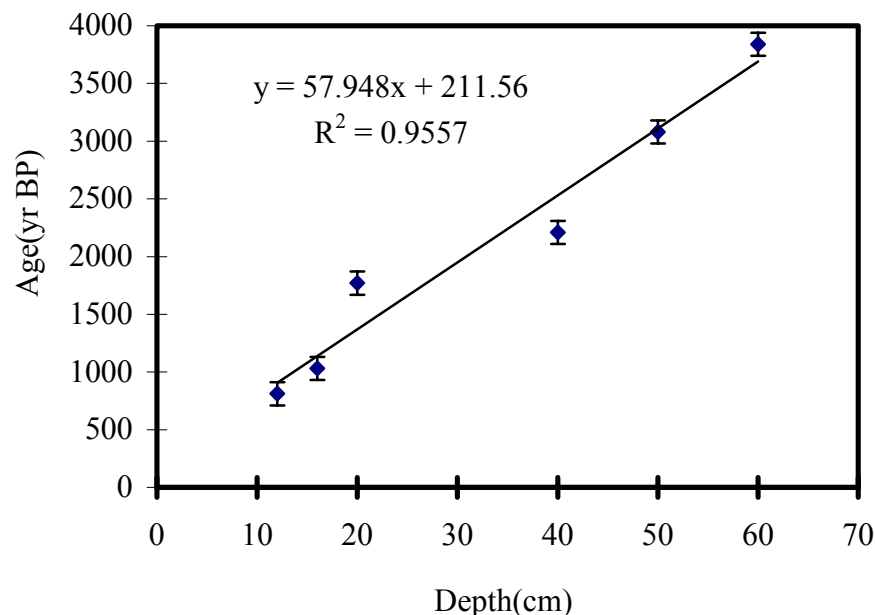
$$y = 57.948x + 211.56$$
$$R^2 = 0.9557$$

Figure 3 Relationship between ^{14}C apparent age and depth of Profile I

Organic Carbon Content and $\delta^{13}C$ Distribution

Figure 2 gives evolution curves for the organic carbon content in the soil of Profile I versus depth and also the $\delta^{13}C$ value versus depth. The 2 curves exhibit similar shaped but opposite trends: a rapid change for the section from the surface to 4 cm in depth, a slow change for the section from 4–10 cm in depth, and slight variation for the section from 10–60 cm in depth. It can thus be inferred that the organic carbon in the soil contains 2 completely different components: one of which has rapidly decomposed with increase in depth, while the other has decomposed extremely slowly. Under the action of microorganisms, the plant residues in the soil are broken down into amino compounds, hydroxybenzenes, saccharides, lignins, etc., and further into humic substances. Organic carbon mainly originates from humic substances, so the distribution characteristics of organic carbon can reflect the accumulation of humic substances in the soil. In general, the soil layer above 10 cm depth is characterized by a relatively loose structure with good air permeability, and the root distribution is well developed with intense biological activity. Therefore, the residual plant material decomposes rapidly, and the dissolved humic substances can be easily removed due to the gley moisture in the surface soil. However, below the 10 cm depth, the physical and chemical properties of the soil and the biogeochemical reactions in the soil are all relatively stable and, consequently, the organic carbon component tends to be stable.

Based on photosynthetic pathways, natural plants can be divided into 3 types: C_3, C_4, and CAM. The $\delta^{13}C$ value for C_3 plants is in the range of –34‰ to –23‰; the $\delta^{13}C$ value for C_4 plants ranges from –22‰ to –6‰; and the $\delta^{13}C$ value for CAM plants is from –20‰ to –10‰ (Deines 1980). The difference between the mean $\delta^{13}C$ values for C_3 and C_4 plants at the same location is about 14‰. Isotope fractionation in organic matter commonly occurs during decomposition and humification processes and the $\delta^{13}C$ value slightly increases, generally by 1~4‰. Organic carbon in soil originates from vegetation and the variation of its $\delta^{13}C$ value reflects the $\delta^{13}C$ variation of the vegetation. The $\delta^{13}C$ values for soil in Profile I ranges from –27.6‰ to –15.7‰. Average $\delta^{13}C$ values for the various soil horizons are –27.4‰ for Layer O, –23.3‰ for Layer A, –18.5‰ for Layer B, and –16.7‰ for Layer C. Based on the $\delta^{13}C$ and ^{14}C apparent ages, the grass and the sparse shrubbery, which cover the slope at the site today, should belong to the C_3 type of vegetation. However, most of the vegetation of this same grassland slope would have belonged to the C_4 type 3000 yr ago.

Atmospheric ^{10}Be and Soil Erosion

Variation curves of ^{10}Be concentration versus depth in Profile I are shown in Figure 4, from which it can be seen that the ^{10}Be concentration distribution has the following characteristics: an exponential increase from 2–30 cm, stability from 30–60 cm, an abrupt 50% increase at 60 cm, and an exponential decrease from 60–150 cm. The total amount of ^{10}Be for the profile can be used to evaluate the retentivity of ^{10}Be in the soils. The ^{14}C apparent age at the boundary of the soil layer is 3840 BP. The age of the weathering pedogenic substances occurring on the rocks is close to or even older than this ^{14}C apparent age. The average global production rate of ^{10}Be is approximately 1.8×10^{-2} atoms $cm^{-2}s^{-1}$ (Beer et al. 1994; Masarik and Beer 1999). On this basis, the gross amount of ^{10}Be accumulated in Profile I should be at least 21.8×10^8 atoms/cm^2. However, ^{10}Be measurements show that the gross amount of ^{10}Be accumulated in the 60-cm-thick soil layer is only 14.4×10^8 atoms/cm^2, the shortfall suggesting a major lateral migration of ^{10}Be. ^{10}Be preservation is closely related to the chemical and mineralogical compositions of the soil grains and the parent materials, and it is generally considered that ^{10}Be can be well retained in soils enriched in clay minerals (Barg et al. 1997). For any given soil profile, ^{10}Be is usually displaced downward together with the migration of organic substances and fine-grained particles. In case of a slope, ^{10}Be can be displaced both physically and chemically. Moreover, it can be displaced transversely toward the valley of the slope due to runoff. In neutral conditions, the distribution coefficient K_d of ^{10}Be between the solid phase (the particle phase) and the liquid phase is as high as 10^5 to 10^6 (Baumgarter et al. 1997). Thus, ^{10}Be is firmly adsorbed on the surfaces of soil particles, and it is probable that the significant lateral migration of ^{10}Be results from soil and ^{10}Be being carried away by underground water, or by the slope wash, in a systematic manner. If so, then the migration of ^{10}Be reflects the status of erosion of this region. This work suggests that about 34% of the pedogenic substances in the area studied have been eroded in the past 3800 yr.

The exponential decrease with depth of the ^{10}Be that is preserved in the interstitial soil of the weathering rocks in the 60–150-cm section suggests that the retention of ^{10}Be is good in this section of soil. It can be seen from Table 1 that the highest ^{10}Be concentration is 0.244×10^8 atoms/g, and the average ^{10}Be concentration is 0.16×10^8 atoms/g at the boundary between the soil and weathering rocks. The ^{10}Be concentration in the upper part of the bedrock is 0.085×10^8 atoms/g. Based on these data, it can be estimated that the 90-cm-thick weathering rock layer above the bedrock has evolved for a period of 1.36 Myr. In this calculation, the average ^{10}Be concentration at the boundary between the soil and the weathering rocks was taken as the initial concentration of the atmospheric ^{10}Be in the 90-cm-thick weathering rock layer. The actual initial concentration of the atmospheric ^{10}Be is difficult to estimate accurately since it is constrained by several factors, such as the precipitation
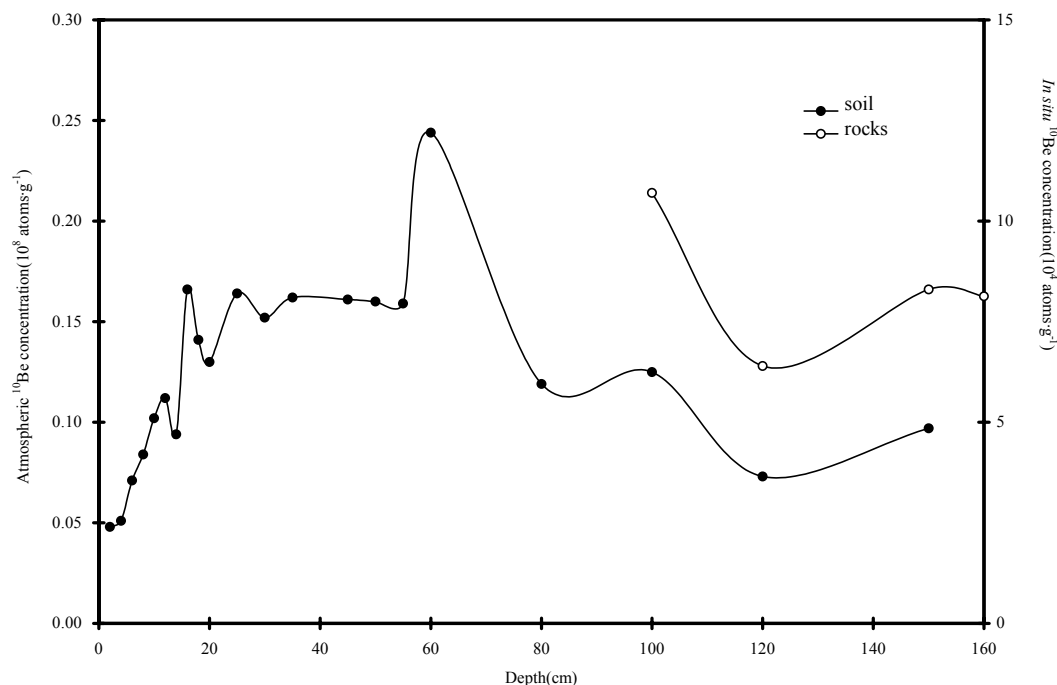
Figure 4 Relationship between ^{10}Be concentration and depth for Profile I. The left ordinate axis and the right ordinate axis represent atmospheric ^{10}Be concentrations in the soil and in situ ^{10}Be concentrations in the rocks, respectively.

flux of ^{10}Be, change of soil erosion rate, ^{10}Be accumulation rate and preservation, etc. Although 1.36 Myr is only a rough estimation, it indicates that the production of soil in this area is rather slow (Heimsath et al. 2000).

In Situ ^{10}Be and Soil Production Rate

The Heshan area (HHLIES-CAS) in Guangdong has a tropical, humid climate which causes intense weathering action. The bedrock in the lower part of the profile is sandy shale, which is overlain by a 90-cm-thick layer of weathering rocks. This layer is comprised mainly of weathered rock fragments of a variety of sizes mixed with interstitial soil. For this work, 3 weathered rock samples were taken at depths of 100 cm, 120 cm, and 150 cm, respectively, together with 1 bedrock sample at 150–160 cm. Quartz was separated from all rock samples and the in situ ^{10}Be concentrations in the quartz were measured by AMS (Table 1 and Figure 4). Concentrations of in situ ^{10}Be vary in the range of $(10.7–6.4) \times 10^4$ atoms/g, which is 2 to 3 orders of magnitude lower than that of the atmospheric ^{10}Be in the same horizon.

Soil in the Heshan Hilly Land comprises mainly accumulated weathering products of the bedrock. In tropical, humid climate zones, dust fallout is negligible. The weathering rate of rock for a stable geomorphic unit is equivalent to the soil production rate (Heimsath et al. 1997). Based on the concentration of the in situ ^{10}Be in rocks, the weathering rate of rocks can be calculated and the relationship between the soil production rate and depth can be derived. Assuming that the soil body is stable with a thickness of h, the soil density is constant, and the bedrock-soil transformation rate is constant, then the ^{10}Be concentration C (atoms/g) of the bedrock at the boundary is the following (Lal 1991):

$$C = P(h, \theta)[\lambda + (\rho_r \epsilon / \wedge)]^{-1}.$$

Here, *P(h, θ)* is the production rate (atoms $g^{-1}y^{-1}$) of ^{10}Be at a depth of *h* and a gradient of *θ* ; $\wedge$ = average attenuation length = 165 g/cm^2; λ = ^{10}Be decay constant = ln2/$T_{1/2}$ = 4.33 × 10^{-7}yr^{-1}; ρ_r = rock bulk density; $\in$ = rate of weathering, which is equivalent to the soil production rate, $\in = (\wedge/\rho_r)[P(h, \theta)/C - \lambda]$.

In order to calculate the rate of weathering $\in$, the following values can be used based on this work: *h* = 60 cm, ρ_r = 2.6 g/cm^3, θ = 18.5°. The production rate *P* of ^{10}Be at the boundary of weathered rock is 1.49 atoms $g^{-1}yr^{-1}$ (Stone 2000). In Table 1, the measured in situ ^{10}Be concentration on the upper part of the weathering rock is 10.7 × 10^4 atoms/g, so the weathering rate $\in$ is estimated to be 8.8 × 10^{-4} cm/yr, which is equivalent to the soil production rate. The cosmic ray exposure time of the upper part of the weathered rock is about 72 kyr. In Table 1, the in situ ^{10}Be concentration in the bedrock is 8.31 × 10^4 atoms/g. After correcting for shielding effects, the production rate *P* of ^{10}Be in the bedrock is 0.36 atoms $g^{-1}yr^{-1}$, yielding a cosmic ray exposure time of about 230 kyr.

As the global ^{10}Be production rate is not known with certainty and its estimates vary by a factor of 2 (Beer et al. 1994; Brown et al. 1989; Kubik et al. 1998; Lal 1991; Monaghan et al. 1985; Pavich et al. 1986; Stone 2000), it is necessary to further reduce the inherent uncertainty regarding the ^{10}Be production rate and improve to reliability of estimating the soil production rate and exposure age.

CONCLUSIONS

A linear relationship between the depth of sample and the ^{14}C apparent age of organic carbon demonstrates that the rock weathering process and the accumulation of organic matter in the slope are relatively stable. The apparent age of the organic carbon at the bottom of Horizon C is 3.8 ± 0.1 kyr.

The ^{13}C value of the grassland slope soil ranges from –27.6‰ to –15.7‰. The grass and the sparsely populated shrubs, which extensively cover the slope in the modern hilly land, all belong to C_3 type vegetation, but most of the vegetation of the grassland slope could have belonged to type C_4 3000 yr ago.

Both ^{14}C and ^{10}Be data indicate that about 34% of the soil substances in the grassland slope have been eroded during the past 3800 yr.

The concentration of in situ ^{10}Be in the weathering rock and the bedrock are 10.7 × 10^4 atoms/g and 8.31 × 10^4 atoms/g, respectively, indicating a weathering rate equivalent to the soil production rate of 8.8 × 10^{-4} cm/yr, and exposure ages of the weathering rock and the bedrock are 72 and 230 kyr, respectively.

ACKNOWLEDGEMENTS

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