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Soil organic matter dynamics after C_3–C_4 vegetation change of red soil in Southern, China: Evidence from natural ^{13}C abundance

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ABSTRACT

Soil samples from natural forests and adjacent farmland were analyzed to investigate the dynamics of soil organic matter of red soil in Southern, China. Based on the $\delta^{13}C$ values and the content of soil organic matter, the data indicated that the turnover of soil organic matter under the virgin forest was slower than that under cultivation. Soil organic matter is fresh in coarse sand and oldest in fine silt and clay. Also, the soil light fraction contained the younger organic matter than soil heavy fraction and bulk soil. Deforestation has accelerated the decomposition rate of soil organic matter and reduced the proportion of active components in SOM and thus soil fertility.

Key words: red soil, carbon dynamics, soil organic matter (SOM), $\delta^{13}C$ values.

Soil organic matter (SOM) is a key source of nutrients for plant production, essential for the maintenance of soil structure and it contributes to the ability of the soil to retain nutrients and water. SOM is the largest terrestrial reservoir of organic carbon with an estimated total content of 1600 Pg C or 80 % of the total terrestrial C storage, exceeding the terrestrial biosphere (600 Pg C) and atmosphere (750 Pg C) storage capacities [1, 2]. Soils can be a source or sink for atmospheric CO_2 depending on land use and management. Therefore, the SOM pool also plays an important role in globe carbon cycle and global warming process [3, 4]. SOM comprises a vast range of different organic structures with a mean residence time in soils ranging from days to millennia [5–7]. An improved understanding of SOM dynamics is central to the development of more environmentally sound and sustainable practices of ecological management [8]. SOM dynamics can be studied in several ways. For example, various chemical fractionation and characterization methods have been commonly used to describe the processes of SOM accumulation and turnover in the last decades [9, 10].

During the past half century, with the rapid development of stable isotope technology and the continuous improvement of the accuracy of instruments, stable carbon isotopes (natural abundances of ^{13}C, expressed as $\delta^{13}C$) has become a powerful tool in ecology, oceanography, geology, environmentology, and biology. The $\delta^{13}C$ value of C_3 plants commonly range from −30 to −22 ‰ (average value − −27 ‰), while the values of C_4 plants range from −15 to −9 ‰ (average value − −13 ‰) [11]. As SOM is an integrated mix of plant litter, roots, and microbial biomass in the soil, so the $\delta^{13}C$ value of SOM reflects the relative proportion of C_3 to C_4 plant material present [12]. In sequential ecosystems, when natural or anthropogenic changes of vegetations from different metabolic pathways (C_3 and C_4) occurred, this large stable carbon isotope difference (> 13 ‰) of C_3 and C_4 plant have proved to be a useful tool for studying SOM dynamics [13−19]. However, the dynamics of SOM are highly variable in different geographic conditions or cultivation modes in different sequential ecosystems all over the world. For example, much farmland is still of high fertility several decades after deforestation in North America [20], whereas SOM is almost exhausted a few years after deforestation in the Brazilian tropics [21].

In the tropical and subtropical regions of southern China there are about 2 million km^2 of red soils, which can be classified as Ultisols in the Soil Taxonomy System of the USA and Acrisols and Ferralsols in the FAO/UNESCO legend. These red soils represent 20 % of the country's total land area, with 520 million people, or 40 % of the nation's population [22]. In the past few decades, due to population pressure and economic development, a vast area of forest has been cleared for crop growth in Southern China. Therefore, land use and land cover transitions (from a C_3 to a C_4 vegetation) provide a good opportunity to utilize the $\delta^{13}C$ to evaluate the SOM dynamics of red soil in Southern, China.

The objectives of this study were: (1) to investigate the effects of vegetation change on SOM; (2) to use the stable carbon isotope method to compare the $\delta^{13}C$ values of SOM in the sequential ecosystems; (3) to describe the dynamics of SOM in different size and density fractions traced by stable carbon isotope.

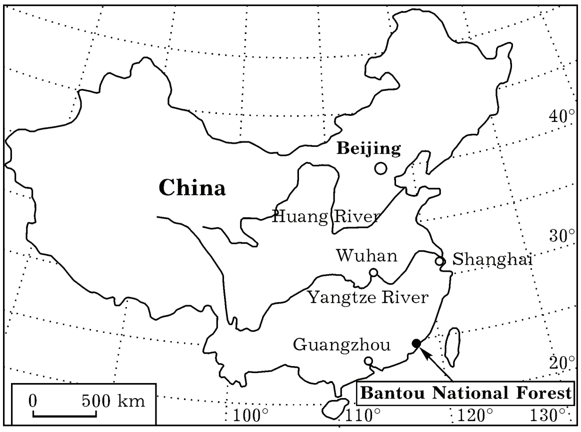

Fig. 1. Geographic location of study site

MATERIAL AND METHODS

Study site. The study was conducted at the boundary area of the Bantou National Forest (24°67′ N, 118°02′ E, 8−21 m a.s.l.), Xiamen, Fujian Province, China (Fig. 1). The mean annual temperature is 21 °C and mean annual precipitation is 1250 mm. We chose 3 sites termed A, B, and C for soil sampling. Each of the sites includes forestland (A_1, B_1, and C_1) and farmland (A_2, B_2, and C_2). The study utilized soil samples from natural forests and adjacent farmland converted to intensive maize cropping in the past several decades (20−30 years). All the soils are predominantly red soil with characteristics as shown in Table 1.

Methods and analysis. At each site, soil samples were collected at depth intervals of 0−5, 5−10, 10−15, 15−20, 20−25, 25−30, 30−

Selected physical and chemical properties of soils analyzed

Sites	Texture Class	Slope, %	pH	Bulk density, g/cm_3	C/N	$\delta^{13}C$ of vegetation, ‰
A_1	Silt loam	3−9	5.9	1.29	11.8	−27.7~−29.3
A_2	Silt loam	<2	5.2	1.36	9.1	−11.3~−11.8
B_1	Loamy Sand	3−6	6.1	1.28	10.6	−27.9~−28.9
B_2	Loamy Sand	<2	4.9	1.26	9.7	−11.3~−11.6
C_1	Sandy loam	2−7	6.3	1.19	12.5	−27.5~−28.4
C_2	Sandy loam	<2	5.2	1.33	10.3	−11.1~−11.9

40, 40–50, and 50–60 cm. All of the samples were air-dried and sieved to 2 mm; any organic matter coarser than 2 mm was treated separately. Three mL of 0,1 M HCl were added per g of soil samples as to minimize any possible contamination by carbonates. Soil and solution were then dried together at 60 °C. The light fraction (LF) and heavy fraction (HF) were separated by flotation in a solution of NaI, with the density adjusted to 1,7 g/cm^3 [23]. Size fractionation was done ultrasonically in a 1 : 10 soil/water suspension, coarse sand (2000–200 μm) and fine sand (200–50 μm) were separated by wet-sieving, coarse silt (50–20 μm) by sedimentation and decantation, fine silt (20–2 μm) by centrifugation, and clay (<2 μm) by flocculation with $CaCl_2$ [15, 24].

Carbon isotope ratios of plant and soil organic samples were measured in terms of production CO_2 by combustion in a sealed quartz tube with CuO at 900 °C. The evolved CO_2 was then purified and analysed on a mass spectrometer (MAT252, Finnigan MAT, USA). Each sample was tested in triplicate; analytical precision on perfectly homogenized samples was ± 0,1 ‰. The results are expressed as $\delta^{13}C$ values:

$$\delta^{13}C\ (‰) = (R_{sample} - R_{reference}) / R_{reference} \cdot 1000 \quad (1)$$

Where R is the ratio of $^{13}C/^{12}C$ and laboratory reference was calibrated against the PeeDee Belemnite (PDB) limestone.

Statistical analysis. All data were examined for homogeneity of variance and normality. Data were analyzed using the univariate general linear model in SPSS 16 for randomized complete block designs. Significantly different main effects were further tested using the least significant differences multiple comparison test [25]. Significant simple effects were tested using the estimated marginal means function in SPSS 16.

RESULTS AND DISCUSSION

Total organic carbon content. Results of total organic carbon content showed marked differences between forestland (A_1, B_1, and C_1) and farmland (A_2, B_2, and C_2) soils (Fig. 2). In farmlands A_2, B_2, and C_2, the total organic carbon content throughout the profiles was <2 %, ranging from 1.87 % at 0–5 cm to 0,36 % at 50–60 cm. On the contrary, total organic

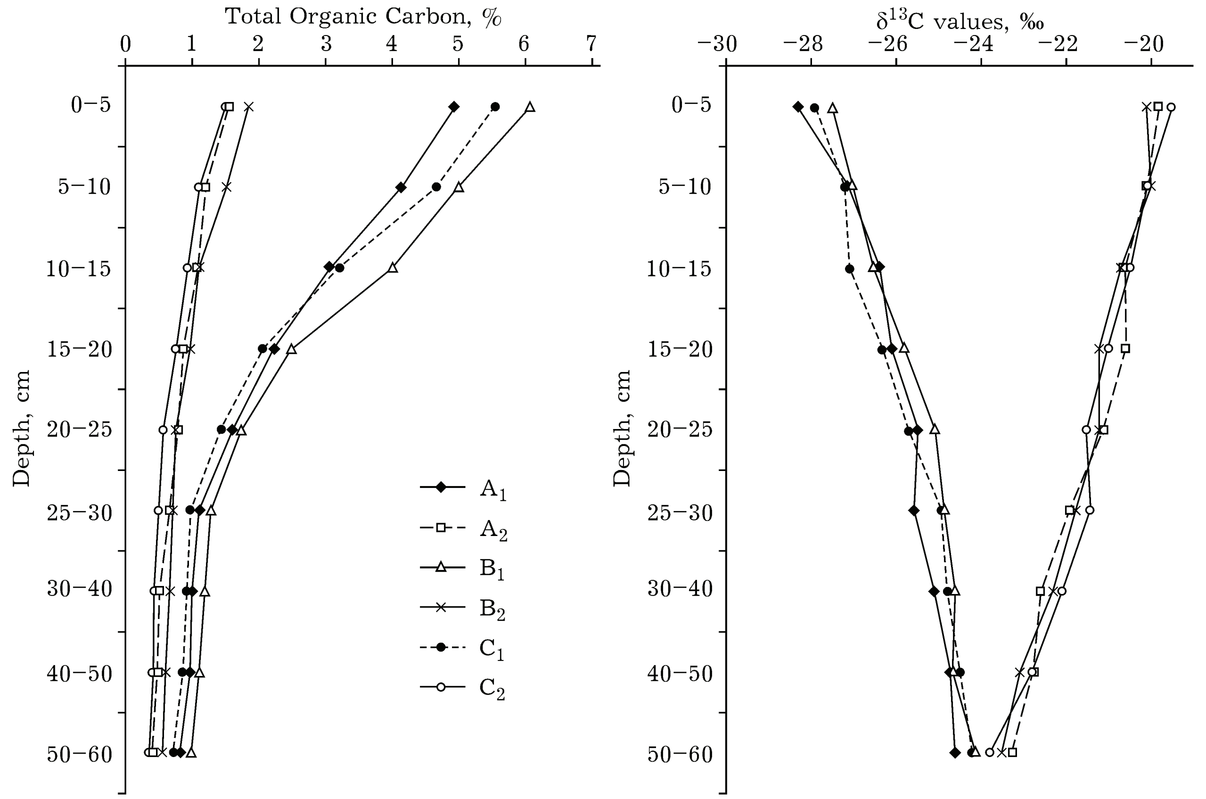

Fig. 2. The total organic carbon content and $\delta^{13}C$ values in the soil profiles

carbon content were consistently higher in forestlands A_1, B_1, and C_1, specifically at the shallow soil layer (0 and 15 cm depth), ranging from 6.11 % at 0–5 cm to 0,75 % at 50–60 cm. The storage of SOM in the soil depends on the balance between additions and losses of C and N [26]. Conventional tillage enhances mineralization of SOM by incorporating crop residue, disrupting soil aggregates and increasing aeration thereby reducing SOM levels. Based on the data, the turnover of SOM under the virgin forest is slower than under cultivation obviously. Thereafter, the content of SOM in the farmland soil would approach a new equilibrium several decades after deforestation. But this equilibrium is established on the basis of low SOM content. Which has also been reported in other studies of red soil in Southern, China [27, 28].

Stable carbon isotopes of SOM. The distribution of SOM $\delta^{13}C$ values on the all study plots is closely tied to the predominant vegetation on each site (see Fig. 2). The forestland (A_1, B_1, and C_1) profiles were typical for soils under C_3 vegetation, where $\delta^{13}C$ values of total soil usually increase with depth. Several rationales can be offered to explain these observed patterns. First of all, the migration and redeposition of clay particles which typically carry a higher ^{13}C content than that of the whole soil, resulting in the increases in $\delta^{13}C$ values with depth [29]. Secondly, differences in decomposition rates of organic compounds with different $\delta^{13}C$ signature. For example, cellulose and hemi-cellulose are often 1–2 ‰ more enriched in ^{13}C, whereas lignin is 2–6 ‰ lower in ^{13}C than in the whole plant tissue [30]. Since lignin is decomposed faster than the cellulose fraction, the isotopic composition of plant residues would change during decomposition. Third, atmospheric $\delta^{13}C$ of CO_2 has decreased by ~1–2 ‰ over the past 130 years [31], as a result of fossil fuel burning and additional biosphere mineralization. In farmland (A_2, B_2, and C_2), the SOM is composed of two types of plant material residues: C_3-derived carbon (SOC_3) from previous forest (approx. equal to $\delta^{13}C$ values of C_3 plants, −28,1 ‰ on average by root, stalk, and leaf measurements in the present study sites) and C_4-derived carbon (SOC_4) from maize crops planted now (approx. equal to $\delta^{13}C$ values of maize, −11,3 ‰ on average by root, stalk, and leaf measurements in the present study sites). Because the discrepancy in $\delta^{13}C$ values between SOC_3 and SOC_4 (> 16 ‰) is greater than the variation in SOC_3 with depth (< 4 ‰) (see Fig. 2), as the percentage of SOC_3 increased with depth, the $\delta^{13}C$ values of SOM in farmland soils clearly decreased with depth profile.

Proportion of SOC_3 and SOC_4 in farmland. In farmland SOM has two isotopically different sources. The proportion of SOC_3 (past residual) and SOC_4 (resulting from conversion to a monoculture of maize) is often calculated using SOM $\delta^{13}C$ values in a simple mass balance mixing formula [13, 15, 17, 32, 33]:

$$\delta^{13}C_{soil} = f \times \delta^{13}C_4 + (1 - f) \times \delta^{13}C_3 \quad (2)$$

$$f = (\delta^{13}C_{soil} - \delta^{13}C_3)/(\delta^{13}C_4 - \delta^{13}C_3) \quad (3)$$

where $\delta^{13}C_{soil}$ = measured soil $\delta^{13}C$ value, $\delta^{13}C_4 = \delta^{13}C$ value for C_4 vegetation, $\delta^{13}C_3 = \delta^{13}C$ value for C_3 vegetation, and f = proportion of SOC_4.

As shown in Fig. 3, the proportion of SOC_4 ranged from 25,6 to 51,2 % with depth, the fraction of SOC_4 in the entire farmland soils were significantly smaller than that of SOC_3. Other studies where cultivation were conducted continuously for several decades after deforestation observed similar results. For exam-

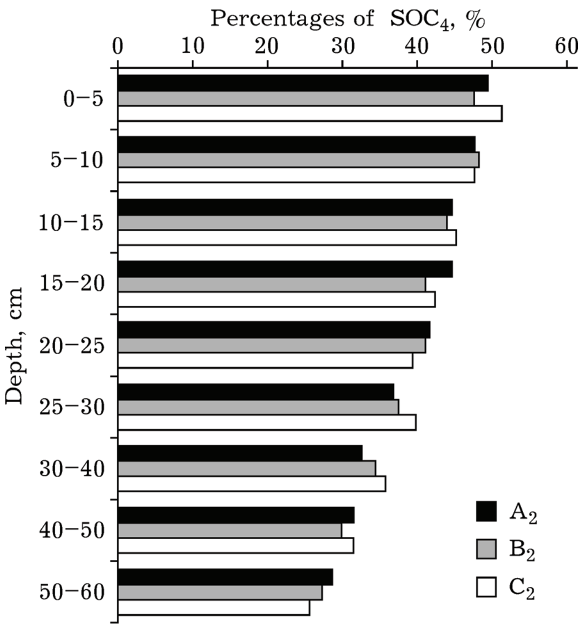

Fig. 3. The percentages (%) of SOC_4 in farmland soil profiles

ple, Wanniarachchi et al. [34] reported that a 29-year old continuous maize crop on an Albic Luvisol in Southern Ontario, had only a 26 % of the total SOC derived from SOC_4 vegetation to a 10 cm depth.

SOM comprises a vast range of different organic structures with a mean residence time ranging from few years of the active fractions to several hundred years of the stable fractions [35]. In their study, the distribution of the active SOM fractions were influenced by soil management factors such as crop rotation, tillage, and fertiliser application. Based on the above total organic carbon content, it is seen that most of the SOM disappeared after deforestation. The remnant component is the stable organic matter, belonging to organo-mineral complexes, which can remain unchanged for a long period of time, but can only maintain the soil structure, not fertility.

Stable carbon isotopes of soil size and density fractions. Various soil fractions, separated in terms of size and density, are differently affected when the levels of whole SOM inputs change [15, 36, 37]. The sand-sized and light fraction SOM are rapidly depleted when forest has been cleared for crop growth, whereas SOM associated with silt and clay are most stable after extended periods of cultivation. A reason for examining SOM size and density fractions is to develop a sensitive indicator that can allow early detection of changes in soil fertility before soil degradation becomes severe.

Stable carbon isotopes of soil size fractions. The $\delta^{13}C$ values of SOM associated with different particle size fractions isolated from the same soil sample differed by as much as 3 ‰ (Fig. 4). In the forestland A_1 site, organic carbon associated with the coarse sand (2000–200 μm) fraction had the lowest $\delta^{13}C$ values, and was generally within 1–3 ‰ of the values for C_3 plant litter and roots. In contrast, fine silt (20–2 μm) and clay (<2 μm) had higher $\delta^{13}C$ values than all other particle size fractions, and were generally within 1–2 ‰ of the values for whole soils. The $\delta^{13}C$ values for fine sand (200–50 μm) and coarse silt (50–20 μm) were usually intermediate between

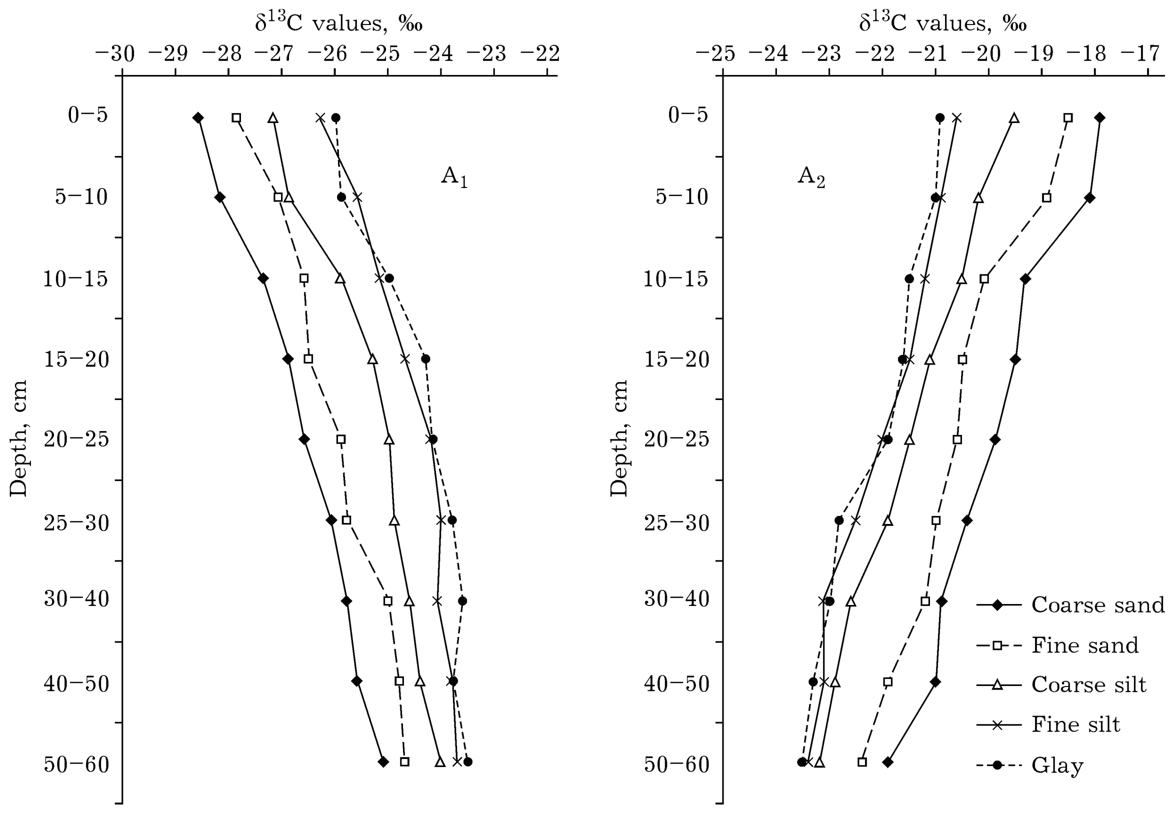

Fig. 4. The $\delta^{13}C$ values of different size fractions in the A_1 and A_2 soil profiles

those of the coarse sand, fine silt and clay fractions. In the farmland A_2 site, SOM has two isotopically different sources. The coarse sand (2000–200 mm) fraction had the highest $\delta^{13}C$ values, In contrast, fine silt (20–2 μm) and clay (<2 μm) had lower $\delta^{13}C$ values than all other particle size fractions, the $\delta^{13}C$ values for fine sand (200–50 μm) and coarse silt (50–20 μm) were usually intermediate between the coarse sand and both fine silt and clay fractions.

In farmland A_2 site, the SOM is composed of two types of plant material residues. The percentage of SOC_4 of different size fractions with depths can be calculated by Equation (3). As shown in Fig. 5, the proportion of SOC_4 of different size fractions decreased with decreasing particle size significantly, ranged from 27,4 to 60,7 % with depth. As SOC_4 inputs were more recent than SOC_3. Therefore, we conclude that the coarse sand contained the youngest organic matter, fine silt and clay contained the oldest organic matter, and the age of fine sand and coarse silt were usually intermediate between those of the coarse sand, fine silt and clay fractions. Organic matter in the sand-size fraction is often more labile than organic matter associated with the clay and silt fractions.

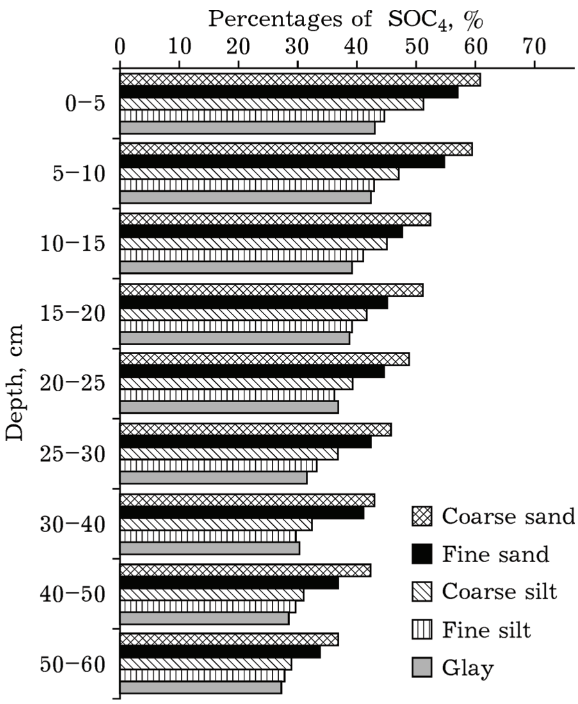

Fig. 5. The percentages (%) of SOC_4 of different size fractions in the A_2 soil profiles

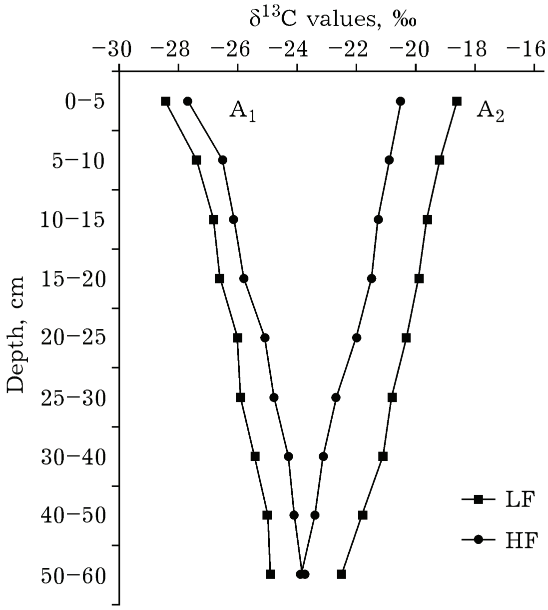

Fig. 6. The $\delta^{13}C$ values of density fractions in the A_1 and A_2 soil profiles

Cambardella and Elliot [38] also reported that the particulate organic matter fraction (POM) < 125 μm was more decomposed than the larger fractions.

Stable carbon isotopes of soil density fractions. As shown in Fig. 6, in the forestland A_1 site, organic carbon associated with the light fraction (LF) had the lower $\delta^{13}C$ values than heavy fraction (HF). In the farmland A_2 site, the rapid loss of native SOM from soils after

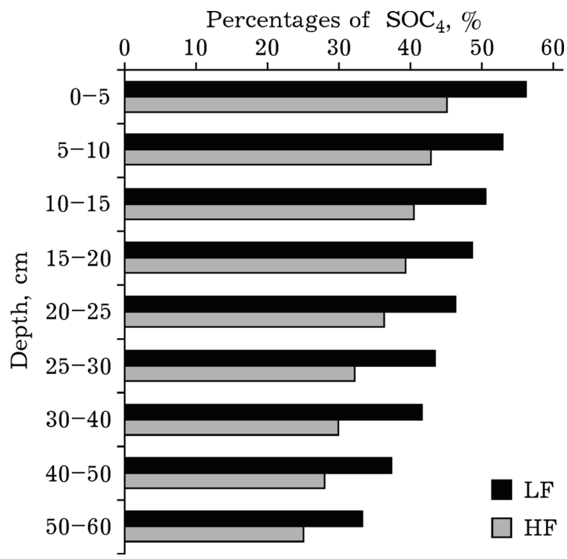

Fig. 7. The percentages (%) of SOC_4 of density fractions in the A_2 soil profiles

forest clearance can be associated with the decomposition of LF materials. So, the $\delta^{13}C$ values of LF higher than the $\delta^{13}C$ values of HF obviously.

In the farmland A_2 site, the proportion of SOC_4 ranged from 25.0 to 56.5 % with depth, and the percentage of SOC_4 in the LF is larger than in the HF (Fig. 7). Therefore, we conclude that the LF contained the younger organic matter than HF and bulk soil as analyzed above. The most organic matter in the HF is stable, belonging to organo-mineral complexes, and the LF has a greater proportion of active components. It can be as an indicator of soil fertility in a forest-cropland vegetation change system.

CONCLUSIONS

From this study, the total organic carbon content was higher in forestland (A_1, B_1, and C_1) than farmland (A_2, B_2, and C_2) obviously, ranging from 6,11 to 0,75 % and from 1,87 to 0,36 %, respectively. The distribution of SOM $\delta^{13}C$ values on all study plots were closely tied to the predominant vegetation on each site. The percentage of SOC_4 in the entire farmland soils was significantly smaller than that of SOC_3. The data has shown that the turnover of SOM under the virgin forest is slower than under cultivation. Deforestation has accelerated the decomposition rate of soil organic matter and reduced the proportion of active components in SOM and thus fertility of red soil in Southern China.

Based on the $\delta^{13}C$ values of SOM and the percentage of SOC_4 associated with different particle size and density fractions, we conclude that the coarse sand contained the youngest organic matter, fine silt and clay contained the oldest organic matter, and the age of fine sand and coarse silt were usually intermediate; and the LF contained the younger organic matter than HF and bulk soil.

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