Science in China Series D: Earth Sciences

Soil organic carbon storage and soil CO_2 flux in the alpine meadow ecosystem

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High-resolution sampling, measurements of organic carbon contents and ^{14}C signatures of selected four soil profiles in the Haibei Station situated on the northeast Tibetan Plateau, and application of ^{14}C tracing technology were conducted in an attempt to investigate the turnover times of soil organic carbon and the soil-CO_2 flux in the alpine meadow ecosystem. The results show that the organic carbon stored in the soils varies from 22.12×10^4 kg C hm^{-2} to 30.75×10^4 kg C hm^{-2} in the alpine meadow ecosystems, with an average of 26.86×10^4 kg C hm^{-2}. Turnover times of organic carbon pools increase with depth from 45 a to 73 a in the surface soil horizon to hundreds of years or millennia or even longer at the deep soil horizons in the alpine meadow ecosystems. The soil-CO_2 flux ranges from 103.24 g C m^{-2} a^{-1} to 254.93 gC m^{-2} a^{-1}, with an average of 191.23 g C m^{-2} a^{-1}. The CO_2 efflux produced from microbial decomposition of organic matter varies from 73.3 g C m^{-2} a^{-1} to 181 g C m^{-2} a^{-1}. More than 30% of total soil organic carbon resides in the active carbon pool and 72.8%—81.23% of total CO_2 emitted from organic matter decomposition results from the topsoil horizon (from 0 cm to 10 cm) for the *Kobresia* meadow. Responding to global warming, the storage, volume of flow and fate of the soil organic carbon in the alpine meadow ecosystem of the Tibetan Plateau will be changed, which needs further research.

Tibetan Plateau, alpine meadow, soil organic carbon, CO_2 flux, ^{14}C signature

The organic carbon inventories stored in the pedosphere is about twice the amount of carbon presented in the atmosphere, and about 3 folds the amount of carbon stored in the vegetation[1]. Soil carbon cycle provides the basis of energy conduction and substance circulation in terrestrial ecosystems, and represents an important component of the global carbon cycle. The CO_2 released from soil respiration is more than ten times that from fossil fuel combustion[2]. The CO_2 efflux from decomposition of soil organic matter (SOM) represents roughly two-thirds of the CO_2 flux between terrestrial ecosystems and the atmosphere[3]. Consequently, any change in the size of the soil carbon pool could potentially alter the atmospheric CO_2 concentration and the global climate. The reservoir of soil carbon should act as a significant source or sink of the atmospheric CO_2 in response to global warming[4]. Tibetan Plateau serves as an important role in the global and Chinese carbon management, ecosystems environmental construction, and startup and regulation of climatic pattern in the Northern Hemisphere owing to its larger surface area (about 1.69% of the global land surface area), the highest altitude, special ecosystems and sensitivity response to global change. Investigators have turned their attention to soil carbon storage and its balance of the Tibetan Plateau ecosystem, but inconsistent results have been obtained because of different data sources and calculating methods[5−17], and there are still very few studies on the soil carbon storage

Received September 11, 2006; accepted January 9, 2007
doi: 10.1007/s11430-007-0055-3
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Supported by the National Natural Science Foundation of China (Grant Nos. 40231015, 40471120 and 40473002) and the Guangdong Provincial Natural Science Foundation of China (Grant No. 06300102)

and release on the basis of the ecosystems function processes[8,10,17]. Up to the present time, no study has yet been devoted to the turnover times of soil organic carbon and CO_2 flux from microbial decomposition of soil organic matter. Investigation of the size and change of soil organic carbon pool in the alpine meadow ecosystem, which is one of major ecosystems in the Tibetan Plateau, is critical for understanding carbon exchanges processes between the soils and the atmosphere, detecting the problems of decreasing soil carbon emitting from human activities and increasing storage and turnover times of soil organic carbon. It is also crucial for understanding the function of the Tibetan Plateau in the global carbon cycle, and the contributions and responses to global environmental change.

In this paper, high-resolution sampling, measurements of organic carbon contents and its radiocarbon (^{14}C) signatures of selected four soil profiles in the Haibei Research Station of Alpine Meadow Ecosystem, Chinese Academy of Sciences (Haibei Station), and application of ^{14}C tracing technology were carried out with the aim of understanding change processes of storage and mean residence times of the soil organic carbon (SOC), and estimating CO_2 released from soil in the alpine meadow on the northeast Tibetan Plateau. The purpose of this study is to provide basic data of soil organic carbon storage, and to forecast the response and feedback of SOC in the alpine meadow ecosystem on the Tibetan Plateau to global warming.

1 Survey of study area

Haibei Station is situated in the northeast Tibetan Plateau, the southern piedmont of Lenglongling Mountain of the eastern Qilian Mountain ranges, and northwest flank of the Datong River valley (between 37º29′ and 37º45′N latitude, and between 101º12′ and 101º33′E longitude) (Figure 1). The elevation of the study area varies from 3200 m to 3600 m, dominated by bottomlands and hills. The mean annual temperature and precipitation are −1.7℃ and about 580 mm, respectively. There is no absolute frost-free period, and the relative frost-free period lasts about 20 d all the year round. The weather of frost, freeze and snowfall (rain and snow mixed) occurs in July. Natural vegetation is alpine meadow. Soil is Cryic Cambisols. Parent material comprises diluvial-alluvial deposits, slope wash-eluvium and fluvioglacial sediments. The soil is almost undisturbed owing to natural grazing in the study area[18].

Alpine meadow (alpine grasses and alpine shrubs) is one of the special natural vegetation formed under the condition of alpine climate after the Tibetan Plateau uplifting in late Cenozoic, which represents a major vegetation type of special plateau zone[19] and mountain vertical zone, occuring widely on the eastern Tibetan Plateau and composing important grassland in China. Alpine shrub comprises broad-leaved *Dasiphora fruticosa* in the northern temperate zone and *Salix oritrepha*, *Sibiraea angustata*, etc., in the Tibetan Plateau. Alpine grass meadow is made up of grassland meadow, real meadow (*Kobresia humilis* meadow, *Kobresia capillifolia* meadow and *Elymus nutans* meadow) and swampy meadow. *Dasiphora fruticosa* shrub meadow and *Kobresia humilis* meadow are the two types of Alpine meadow widely distributed in the Haibei Station. *Dasiphora fruticosa* shrub meadow occurs mostly on the mountain slope non-lighted, humid bottomland, piedmont diluvial fan, low river terrace and mountain slope lighted above 3800 m elevation. *Kobresia humilis* meadow is situated primarily on the bottomland, wide valley, foot of mountain and south-facing slopes[20]. Three soil types under natural alpine meadow vegetation (*Dasiphora fruticosa* shrub, *Kobresia humilis* at bottomland, *Kobresia humilis* at slow-slope) and the other under planted *Arrhenatherum elatius* since 1978 (cultivating 20–30 cm depth in spring, seeding on the last ten-day of May, fertilizing carbamide and dung of sheep and cattle, harvesting on the middle ten-day of September) were selected and sampled in the Haibei Station in July 2003 in an attempt to investigate the influences of vegetation, landforms, and human activities on soil carbon cycle of alpine meadow (Figure 1). Some important characteristics of study sites are summarized in Table 1.

2 Materials and methods

2.1 Sampling

A sampling pole was placed on a side of the formerly excavated exploratory trench (100 cm×80 cm×150 cm) in the selected field, where aboveground plants and litters were removed from the sampling pole surface (50 cm ×50 cm). Soil samples of 1.5－2.0 kg, by the way of vertically, successively and completely thin-layer sampling, were taken at different intervals with stainless steel knife and shovel. For the four profiles, soil samples were taken successively from ground down to 30 cm

Table 1 Major characteristics of the study area (data from refs. [18, 21])

Ecosystem types	Location	Altitude (m)	Land type	Biomass above ground (kg m^{-2} a^{-1})	Land use
Kobresia humilis at bottom-land	37°37′N, 101°19′E	3220	bottomland	0.35	light free-grazing in winter and spring
Kobresia humilis at slow-slope		3230	slow-slope gradient 5°–7°	–	
Dasiphora fruticosa shrub	37°40′N, 101°19′E	3352	bottomland	0.27	
Arrhenatherum elatius	37°37′N, 101°19′E	3220	bottomland	2.31 (fresh)	mowing

depth at 2 cm intervals, and samples were further taken at 5 cm, 10 cm and 20 cm intervals at 30—60 cm, 60—80 cm and 80—100 cm respectively.

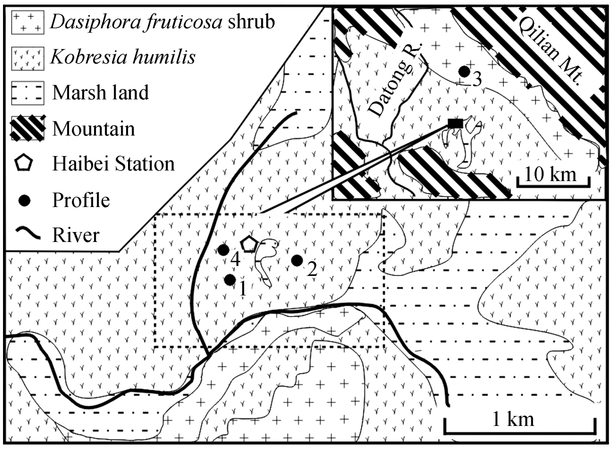

Figure 1 Schematic diagram of the studied profiles sites. 1, *Kobresia humilis* meadow Mattic Cryic Cambisols site at bottomland (ASC I); 2, *Kobresia humilis* meadow Mattic Cryic Cambisols site at slow-slope (ASC II); 3, *Dasiphora fruticosa* shrub meadow Mollic Cryic Cambisols site (JLM) and 4, *Arrhenatherum elatius* site (YMC).

2.2 Laboratory methods

(i) Measurement of soil organic carbon content. Air-dried soil samples were carefully homogenized and sieved to pass 1 mm. Litter and root materials were separated manually from the soil. Followed by weighing up some 10—200 mg of soil samples by electronic balance (AB104-N, Produced by Mettler-Toledo Group, d = 0.0001 g), dispersing with distilled water and treating with 10% HCl to remove any carbonate. The samples were oven dried at 80℃. CO_2 was produced by combustion of the sample with CuO and silver thread under vacuum at 800°C for 15 min and purified cryogenically. Weight percentage of C in a sample was determined from the CO_2 yield[22,23]. Carbon contents of soil samples, C (%), were converted to soil organic carbon (SOC) content (kg hm^{-2}) for a single layer:

$$\mathrm{SOC}_i = C_i \times h_i \times \rho \times 10^4, \tag{1}$$

where C_i represents carbon contents of soil samples, %; h_i represents soil layer thickness, m; ρ is the mean bulk density, g cm^{-3}; i denotes soil layer number.

The value of mean bulk density of sub-alpine meadow soil, 1.20 g cm^{-3}, is selected in the course of calculating due to the absence of field measured value of soil bulk density in the study area[14].

SOC storage of any ecosystem in the study area, SOC_t (kg hm^{-2}), is calculated as

$$\mathrm{SOC}_t = \sum_{i=1}^{k} \mathrm{SOC}_i \ (k = 1, 2, 3\ldots\ldots 24). \tag{2}$$

(ii) Measurement of ^{14}C radioactivity of soil organic matter. Air-dried soil samples (110–500 g)，removing litter and root materials, were dispersed with distilled water and treated with 10% HCl to remove any carbonate, followed by washed repeatedly with distilled water until they were neutral and oven dried at 105°C, then a series of reaction were conducted under vacuum.

$$\text{Soil sample} + O_2 \xrightarrow[800°C]{\text{combustion}} CO_2$$

$$10Li + 2CO_2 = 4Li_2O + Li_2C_2$$

$$Li_2C_2 + 2H_2O = C_2H_2 + 2LiOH$$

Liquid benzene was polymerized by activation of the CrO_3-Al_2O_3-SiO_2 activator.

$$3C_2H_3 \xrightleftharpoons{\text{activator}} C_6H_6$$

After about 30 days, ^{14}C activity of the synthesized C_6H_6 was measured on a 1220 Quantulus ultralow level liquid scintillation spectrometer produced by WALLAC Ltd., and the analytical precision for ^{14}C analysis is better than 1%.

The ^{14}C content of soil organic matter (SOM) reflects the difference between input rate of carbon and turnover rate of SOM. ^{14}C data are reported as $\Delta^{14}C$. Negative $\Delta^{14}C$ values indicate that, on the average, the carbon has resided in the soil long enough to reflect radioactive decay of cosmogenic ^{14}C. Positive $\Delta^{14}C$ values show significant amounts of “bomb” ^{14}C[24]. $\Delta^{14}C$ values can be calculated according to eq. (3):

$$\Delta^{14}\mathrm{C}=\left(\frac{A_{\mathrm{SN}}}{A_{\mathrm{abs}}}-1\right)\times 1000\,‰, \qquad (3)$$

where A_{SN} is the normalized sample ^{14}C activity, A_{abs} is the absolute international standard activity, which is 0.95 times the activity, in 1950, of the NBS oxalic acid (SRM-4990) normalized to $\delta^{13}C = -19‰$ with respect to PDB[25].

The SOC and ^{14}C analyses were carried out in the Key Laboratory of Isotope Geochronology and Geochemistry, Guangzhou Institute of Geochemistry, Chinese Academy of Sciences.

(iii) Estimate of soil CO_2 flux. Exchange ^{14}C between living plants and the atmosphere through photosynthesis, which makes living plants have some the same ^{14}C content as the contemporary atmosphere. Plants no longer absorb ^{14}C from the atmosphere when they die, and their ^{14}C contents decreases due to radioactive decay. Hence, ^{14}C contents of organic matter can serve as a "clock" to estimate the mean residence time of SOC and the CO_2 flux released from the decomposition of soil organic matter on different time scales.

^{14}C-dating was applied to studying soil humus in the late 1960s[26], and Carbon isotopes tracing provides a very powerful tool for studying soil carbon cycle originating from the late 1970s[27]. Technology of ^{14}C tracing can be used to study soil carbon dynamics on time scales of years or decades to hundreds or even thousands of years, especially, ^{14}C produced from atmospheric nuclear test ("bomb " ^{14}C) during the period from 1954 to 1962 is a useful tracer to study soil organic matter (SOM) turnover[24,27–29]. Shortly after the atmospheric nuclear test ban treaty enacted in 1963, the $^{14}CO_2$ concentration in the Northern Hemisphere was twice as high as the natural equilibrium value. $\Delta^{14}C$ value has been steadily declining toward the prebomb level thereafter due to equilibration by the oceans and the terrestrial ecosystems, and dilution by "^{14}C-dead" CO_2 from fossil fuel combustion[24,30,31]. Up to now, $\Delta^{14}C$ value in the atmospheric CO_2 is higher than that of the prebomb[32].

^{14}C content of SOM is restricted to three processes: (1) ^{14}C is absorbed from the atmosphere, (2) ^{14}C is released from SOM by microbial decomposition, and (3) radioactive decay of ^{14}C. ^{14}C-dating is invalid due to the values of ^{14}C more than zero at the upper of soil profiles by influence of "bomb" ^{14}C. To quantify turnover times for SOM, the mathematical model suggested by Cherkinsky and Brovkin (1993) was used. The model is expressed as[33]

$$A(t)=A(t-1)-(m+\lambda)\times A(t-1)+mA_0(t), \qquad (4)$$

$$\frac{A(1955)}{A_{\mathrm{abs}}}=\frac{m}{m+\lambda}, \qquad (5)$$

where $A(t)$ is the ^{14}C radioactivity of SOM in the year of sampling (t >1955); $A(t-1)$ is the ^{14}C radioactivity of SOM in the preceding year of sampling; $A_o(t)$ is the ^{14}C radioactivity of atmosphere in the year of sampling; A_{abs} is the ^{14}C radioactivity of the absolute international standard; m(a^{-1}) is the turnover rate of SOM; λ (a^{-1}) is the decay constant of ^{14}C, λ= 1/8033, and A (1955) is the ^{14}C radioactivity of SOM in 1955.

As far as field observed litters and roots composition, and natural environmental characteristics in the study area are concerned, we assume that ^{14}C absorbed by plants from the atmosphere would be incorporated into the soil in the next year, hence, $A_o(t-1)$ took the place of $A_o(t)$ in eq. (4) in the course of calculating.

Variation of ^{14}C content in the atmosphere since 1977 is derived from the $\Delta^{14}C$ values of the atmospheric CO_2 reported by Levin and Kromer as follows[32]:

$$Y = 354.03\exp^{-0.0614(t-1976)}\ (R^2=0.9958,\ n=20), \qquad (6)$$

where Y is the $\Delta^{14}C$ value of the atmospheric CO_2 in t year (t >1977) (‰). The atmospheric $\Delta^{14}C$ value since 1997 has been calculated by eq. (6). Then, the specific activity of the atmospheric ^{14}C in t year can be derived from the definition of $\Delta^{14}C$ and pM.

In this study, the ^{14}C specific activity of SOM in the preceding year of sampling (i.e. 2002) was calculated by eq. (4). It is assumed that the amount of ^{14}C in soil is stabilized by 1955, first, the value of m was selected, A(1955) was calculated by eq. (5); secondly, A(1955) and the selected m were put into eq. (4), the ^{14}C radioactivity of SOM in the preceding year of sampling (2002) was calculated by changing m value time after time. The value of m was the turnover rate of the SOM until the value of A(2002) calculated was close to the measured value of the ^{14}C radioactivity of the soil sample with calculating precision of 0.00002 a^{-1}.

For the soil layers with the ^{14}C values of the SOM less than zero below the certainty depth of the soil profiles, the influence of "bomb" ^{14}C was negligible. It is assumed that these soil layers were stable and closed, and hence there was no soil mixing and the decomposition of SOM is isolated. According to eq. (5) and the definition of $\Delta^{14}C$, the turnover rate of the SOM in these soil layers can be calculated as[34]

$$m = -\lambda\left(1+\frac{1000}{\Delta^{14}C}\right), \quad (7)$$

where λ is the decay constant of ^{14}C, λ = 1/8033.

The turnover time (a) of the SOM can be calculated as

$$t = \frac{1}{m}. \quad (8)$$

CO_2 production released from microbial decomposition of SOM of a single soil layer can be calculated as

$$F_i = \rho \times h \times C \times m, \quad (9)$$

where F_i is CO_2 production of a single soil layer, g C cm^{-2} a^{-1}; h_i represents soil layer thickness, m; C_i represents carbon content of soil samples, %; ρ is mean bulk density, g cm^{-3}; m is the turnover rate of the SOM, a^{-1}; i denotes soil layer number.

The CO_2 flux released from microbial decomposition of SOM of such ecosystem in the study area is expressed as

$$F_t = \sum_{i=1}^{k} F_i \quad (k=1,2,3\ldots.24), \quad (10)$$

where F_t is the CO_2 flux from decomposition of SOM in the given ecosystem, g C m^{-2} a^{-1}.

3 Results and discussion

3.1 Storage of soil organic carbon in different ecosystems

The difference of SOC content and its decrease with depth are obvious in different alpine meadow ecosystems in the study area (Table 2; Figure 2). The SOC content of the *Dasiphora fruticosa* shrub is the highest among all of the studied alpine meadow ecosystems. The SOC stored in the surface soil horizons (0–10 cm depth) of the three natural alpine meadow ecosystems (ASC I, ASC II and JLM) constitutes 7.07×10^4 kg C hm^{-2} or 32% of the total organic carbon (from 0 to100 cm depth), 8.16×10^4 kg C hm^{-2} or 34%, and 8.59×10^4 kg C hm^{-2} or 28%, respectively, and decreases remarkably with depth. However, there is only 5.29×10^4 kg C hm^{-2} or 17% of the total organic carbon (from 0 to100 cm depth) at the corresponding depth for the *Kobresia humilis* site (YMC). Some 25.18% of SOC stored at the surface soil layers (0—10 cm depth) is lost due to cultivation. The SOC contents of ASC I below 10 cm depth are significantly lower than those of YMC, which results from the obvious downward shift of the SOC by converting shallow-rooted *Kobresia humilis* into deep-rooted *Arrhenatherum elatius*, which is consistent with root distributions in the soil profiles. At the *Kobresia humilis* site, plenty of roots in the surface horizons (0—10 cm depth), comprise 67.29%—69.7% of the total roots biomass (from 0 to 100 cm depth), which drop drastically below 10 cm depth. At the *Arrhenatherum elatius* site, merely 43% of the total roots biomass (from 0 to 100 cm depth) is present in the surface soil horizons (0—10 cm depth), but the roots are more abundant at 10—55 cm depth.

Table 2 Content of soil organic carbon in different alpine meadow ecosystems

Depth (cm)	Content of soil organic carbon (×10^4kg C hm^{-2})			
	ASC I	ASC II	JLM	YMC
0–2	2.45	1.80	2.57	1.06
2–4	1.51	1.79	1.80	1.04
4–6	1.28	1.58	1.70	0.92
6–8	0.97	1.42	1.18	0.99
8–10	0.85	1.57	1.33	1.28
10–12	0.78	1.34	1.31	1.10
12–14	0.69	1.16	1.10	0.95
14–16	0.66	1.04	1.04	1.18
16–18	0.40	1.00	0.87	1.15
18–20	0.75	0.85	0.87	1.24
20–22	0.67	0.91	0.81	1.20
22–24	0.58	0.68	0.78	1.04
24–26	0.61	0.62	0.75	1.03
26–28	0.63	0.59	0.70	0.90
28–30	0.58	0.53	0.67	0.89
30–35	1.30	1.25	1.93	1.92
35–40	1.07	1.00	1.46	1.91
40–45	0.94	1.01	1.31	1.97
45–50	0.91	0.89	1.26	1.44
50–55	0.77	1.05	1.13	1.24
55–60	0.66	0.32	1.10	1.09
60–70	1.16	0.57	1.77	1.89
70–80	0.70	0.39	1.54	1.42
80–100	1.18	0.95	1.76	1.41

SOC content of the study area is expressed as SOM content and calculated in terms of SOM content multiplying by Bemmelen coefficient (i.e., 0.58)[35,36]. Notice that there is difference between the SOC contents of the ecosystems investigated in this paper and that investigated by other researchers, which is caused by the approaches used by individual authors and the depths studied (Table 3).

The difference in SOC storage (SOC_t) is evident among the investigated ecosystems (Figure 3). The sequence of SOC_t value, in turn, is 30.75×10^4 kg C hm^{-2} for JLM, 30.24×10^4 kg C hm^{-2} for YMC, 24.32×10^4 kg C hm^{-2} for ASC II and 22.12×10^4 kg C hm^{-2} for ASC I, respectively, with an average of 26.86×10^4 kg C hm^{-2} in the study area.

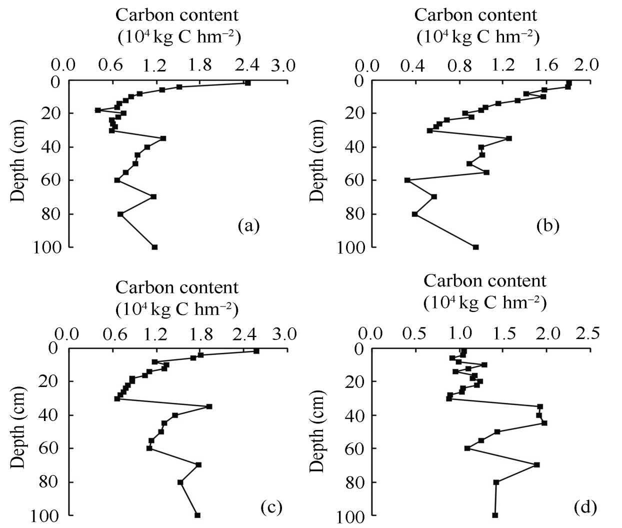

Figure 2 SOC variation pattern with depth of different ecosystems in the study area. (a) ASC I profile; (b) ASC II profile; (c) JLM profile; (d) YMC profile.

Table 3 Comparison of soil organic carbon contents in the study area

Profile	Depth (cm) 0–4(10)	4(10)–24(20)	24(20)–45(40)	45–70	Reference
	C (g kg^{-1})				
ASC	82.67	31.75	21.87	12.19	this paper
	(50.17)	(45.37)	(29.67)		[35]
	(61.13)		(27.67)		[36]
	101.00	40.20	20.40	5.60	[37]
JLM	91.19	42.61	24.29	18.27	this paper
	(119.65)		(74.36)		[36]
Alpine meadow soil	70.30 (no depth ascertained)				[38]

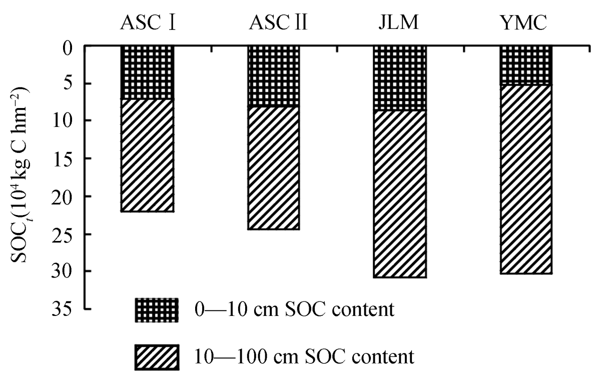

Figure 3 Comparison of SOC_t (1.0 m) in different ecosystems in the study area.

SOC stems mostly from debris, excreta and secretion of plant, animal and microbe. Changes in soil organic carbon storage among different ecosystems are primarily controlled by primary production input process and decomposition process. To some extent, decay rates of plant litters and roots, and the process of forming and decomposing of SOC are determined by the microclimatic condition in the surface soil horizon and deeper soil layer influenced by vegetation types, which causes the difference in SOC storage among different ecosystems and different soil layers[39]. Higher SOC storage under the *Dasiphora fruticosa* shrub meadow results from the combined effects of plenty of roots biomass (27947 kg hm^{-2}) and lower soil temperature (0—20 cm depth, mean annual soil temperature 1.83℃), higher soil moisture (mean annual 30%) coupled with longer freezing time resulting in slow-decomposing SOC; in contrast, lower roots biomass (25745 kg hm^{-2}) and fast-decomposing SOC originated mostly from the combined effects of higher soil temperature (0—20 cm depth, mean annual soil temperature 3.15℃), lower soil moisture (mean annual 28%) coupled with shorter freezing time[18], which cause sufficiently lower SOC storage under the *Kobresia humilis* meadow than that under the *Dasiphora fruticosa* shrub meadow. Compared with the *Kobresia humilis* meadow, the low SOC content of the surface soil in the *Arrhenatherum elatius* site is caused by the reducing accumulation of litters and roots, shifting SOM downwards into the deep layers and fast decomposing of SOM in the surface soil layers, as a result of the soil temperature fast rising, soil moisture dropping, and aerate condition meliorating owing to cultivating activities. The SOC_t values of the deeper soil layers in the *Arrhenatherum elatius* site are significantly higher than those under the *Kobresia humilis* meadow at bottomland, resulting from increasing biomass inputs in the soil layers (below 10 cm) as a result of the combined effects of larger roots biomass, slow-decomposing SOC, and embedding undecomposed debris and cattle dung in the subsoil layers.

The estimation of SOC storage in the study area is evidently lower than those (36.50 kg C m^{-2} (82.9 cm)[11], 53.13 kg C m^{-2} (Qinghai Province, 65 cm)[12] and 50.25 kg C m^{-2} (80 cm)[14] respectively) on Cryic Cambisols obtained by the other investigators, which is likely to be attributed to remarkable difference of SOC storage in the surface soils among subgroup Cryic Cambisols (e.g., SOC storage of the surface soils represents 58－87 g kg^{-1} and 58－116 g kg^{-1} in Mat-Cryic Cambisols and Mol-Cryic Cambisols respectively[37]), unequal numbers of soil profile, non-uniform soil depths and different calculating methods.

3.2 Comparison of SOC storage

By comparison, the SOC_t value of the study area is significantly higher than that of Inner Mongolia grassland and that of planted pasture in the tropic zone, and is also remarkably higher than that of the forest ecosystem located in different climatic zones[23,40,41] (Figure 4), which is explained by means of vast accumulation of SOM resulted from advantaged environmental condition of physical geography, i.e. high altitude, lower mean annual soil temperature, higher soil moisture and longer soil freezing time in the study area. In contrast, warmer and drier environment in the Inner Mongolia grassland and hotter and wetter environment in the tropic zone are apt to the SOC decomposition. It is evident that alpine environment on the Tibetan Plateau is in favor of the SOM accumulation. Therefore, it is concluded that the SOC storage is the comprehensive effects of geographic location, altitude and vegetation type. The SOC pool of the alpine meadow ecosystem in the Tibetan Plateau plays an important role in Chinese or even in the global carbon budget[11].

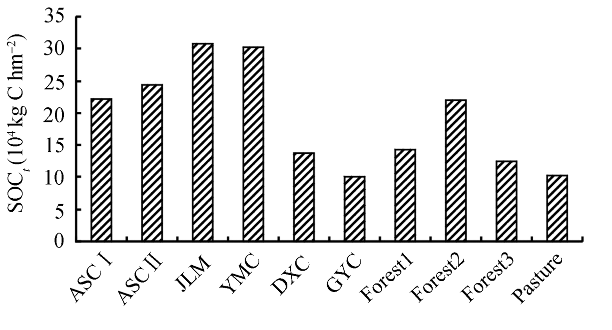

Figure 4 Storage comparison of SOC in different ecosystems located at different climatic zones. DXC (43°00′N, 117°29′E) and GYC (43°34′N, 116°40′E) are situated in Inner Mongolia grassland, respectively; forest 1 is situated in the sub-alpine woodland of the Mt. Gongga in eastern Tibetan Plateau[40]; forest 2 (45°24′N, 127°40′E) is situated in northeast China[41]; forest 3 and pasture are situated in the Atlantic Zone of Costa Rica, pasture is planted about 25 years earlier[23].

3.3 Turnover time of soil organic matter

Pronounced difference exists in ^{14}C contents of soil organic matter (SOM) and their depth-curve among the different ecosystems in the study area (Figure 5). The $\Delta^{14}C$ values of SOM in the topsoil horizons (0－4 cm) are more than zero, and the $\Delta^{14}C$ values of ASC I profile and ASC II profile (108‰－148‰) are more positive than those of JLM profile and YMC profile (3‰－40‰), implies that penetrating depth of the "bomb" ^{14}C was about 4 cm. The $\Delta^{14}C$ values of SOM (below 10 cm) are less than zero, and the $\Delta^{14}C$ values of JLM profile are larger than those of ASC I profile. This phenomenon is explained by the *Dasiphora fruticosa* shrub meadow mostly consisting of deep-rooted plants, increasing carbon inputs from more roots biomass at subsoil horizons (below 10 cm). However, at the *Arrhenatherum elatius* site, organic matter in the topsoil horizons (0－10 cm) is relatively depleted in ^{14}C, and in subsoil layers (10－55 cm) it becomes relatively enriched in ^{14}C, and the $\Delta^{14}C$ values are larger than that in the *Kobresia humilis* meadow at bottomland, which implies increasing young carbon inputs in these subsoil layers. This phenomenon is caused by the fewer young carbon inputs and its faster decomposition in the topsoil horizons and, mixed effect of "old-carbon" from the deeper soil layers by cultivation activity. From 10 to 55 cm, however, the ^{14}C enrichment results from the combined effects of increasing young carbon inputs from more roots and, in part, downward translocation of young carbon from the surface horizons in the forms of dissolved or particulate organic carbon[22].

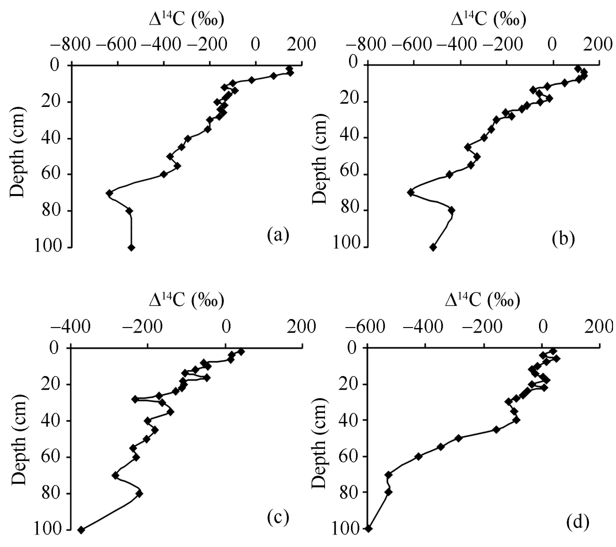

Figure 5 ^{14}C contents of soil organic matter in different ecosystems in the study area. (a) ASC I profile; (b) ASC II profile; (c) JLM profile; (d) YMC profile.

The SOM with faster turnover rates has more positive $\Delta^{14}C$ values, the SOM with slower turnover rates or passive organic matter has more negative $\Delta^{14}C$ values[24]. On the basis of these criteria, the SOM (0—4 cm) under the *Kobresia humilis* meadow has the shortest turnover time, and the SOM under the *Dasiphora fruticosa* shrub meadow and the *Arrhenatherum elatius* site reside the longest time at the corresponding depth. Turnover times of SOM in the study area were estimated by means of eq. (8) (Table 4; Figure 6).

Table 4 Turnover times of soil organic carbon in the Alpine meadow

Depth (cm)	Turnover times (a)			
	ASC I	ASC II	JLM	YMC
0–2	47	73	171	171
2–4	45	53	239	294
4–6	102	54	246	147
6–8	167	71	474	235
8–10	898	149	386	129
10–12	1275	221	682	272
12–14	791	738	935	181
14–16	1081	494	403	295
16–18	1209	131	975	247
18–20	1602	476	973	300
20–22	1256	1025	1015	278
22–24	1465	1247	1194	418
24–26	1319	2066	1654	561
26–28	1517	1773	2450	784
28–30	2031	2626	1582	1060
30–35	2102	2895	1337	854
35–40	3400	3382	2012	772
40–45	3866	4656	1786	1526
45–50	4759	3927	2037	3339
50–55	4191	4437	2522	4421
55–60	5319	6464	2407	6050
60–70	14155	12780	3204	9231
70–80	9860	6311	2283	9290
80–100	9527	8695	4768	12221

The organic carbon turnover rates in the topsoil horizons (from 0 to 4 cm) are faster, with turnover times of 45—73 a under the *Kobresia humilis* meadow, and more than 100 a (171—239 a) in the corresponding depth under the *Dasiphora fruticosa* shrub meadow (Table 4). Downwards, the SOC turnover times increase with depth from hundreds of years to millennia or even longer in the study area. The shorter turnover times of organic carbon in the topsoil horizons under the *Kobresia humilis* meadow imply that these SOC pools will respond sensitively to changes in land use or climate.

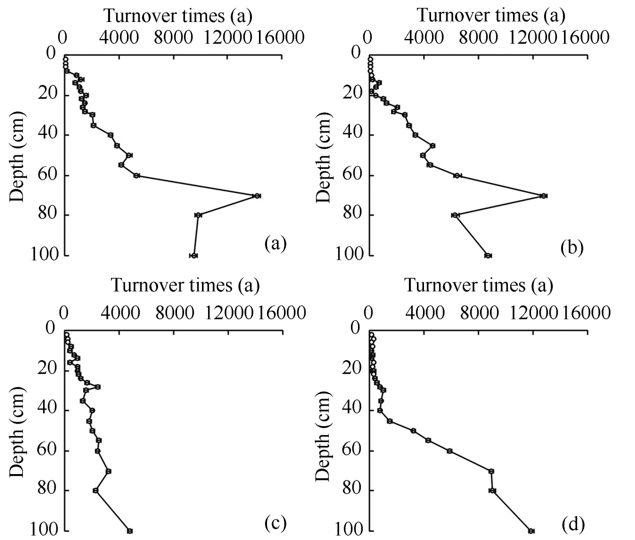

Figure 6 The turnover times of SOM of different ecosystems in the study area. (a) ASC I profile; (b) ASC II profile; (c) JLM profile; (d) YMC profile

The SOC turnover times (171—294 a) in the topsoil horizons under the *Arrhenatherum elatius* are longer than those for the natural meadow ecosystems, which is consistent with the result that the model-estimated SOC turnover times in the agricultural soil are longer than those in the natural grassland[22]. Longer SOC turnover times result from higher inert fraction, stemming from combined effects of diluting by "old-carbon" from the deeper soil layers, decreasing of litters and quick decomposing SOM, in part absorbed by crops, in part lost by leakage, in the topsoil horizons under the *Arrhenatherum elatius* due to cultivation activity.

Soil organic matter is a complex system comprising components of different turnover times. Generally, SOM is at least divided into three carbon pools (i.e. "active" pool, "slow" pool and "passive" pool) or fast-cycling C pool (active pool and slow pool) and passive pool[4] or "active" pool and "passive" pool[22] in terms of SOM turnover times in the soil carbon cycle study. Evident component difference of SOC pools occurs in different ecosystems, resulting from different vegetation, soil structure, soil microclimate (soil temperature and soil moisture), landform, and way and strength of land-use. Relatively "active" pool (t < 100 a) represents 8% of the total SOC pool (from 0 to 100 cm depth) for the *Kobresia humilis* meadow at slow-slope and 4% for the *Kobresia humilis* meadow at bottomland. SOC primarily consists of "slow" carbon (100 a < t < 1000 a) and "passive" carbon (t > 1000 a), with absence of relatively "active" carbon for the *Dasiphora fruticosa* shrub meadow and *Arrhenatherum elatius*.

Turnover times of fast-cycling SOC vary with climate and vegetation[4]. The SOC turnover times (45—73 a) in the topsoil horizons of the study area are significantly longer than that (7.4 a) for the natural grassland on the western slope of the Sierra Nevada Mountain Range in central California[22]. They are also larger than that (1.6—7 a) for the forest ecosystem in the subtropics in China[34] and the average turnover times (22 a[42]/32a[43]) of the world. This phenomenon results from slow SOM decomposition owing to alpine eco-environment that results in low soil temperature and weak microbe activity. The alpine eco-environment is one of the primary factors of considerable SOC storage in the alpine meadow ecosystem on the Tibetan Plateau.

3.4 Soil CO_2 flux released from decomposition of organic matter

Soil respiration is the only way of CO_2 released from the pedosphere to the atmosphere and one of primary sources for the atmospheric CO_2. Soil respiration is defined as the total CO_2 yields in intact soils resulting from the respiration of soil organisms (microbial decomposition of organic matter), roots and mycorrhizae, and chemical oxidation of carbon-containing materials[43], of which biological processes represents dominant[44]. For the study area, the soil-CO_2 flux from microbial decomposition of organic matter estimated by means of eq. (10), ranges from 73.3 to 181 g C m^{-2} a^{-1}. The soil-CO_2 flux from microbial decomposition of organic matter (F_t) is, in turn, 181 g C m^{-2} a^{-1} for the *Kobresia humilis* meadow at bottomland, 177.8 g C m^{-2} a^{-1} for the *Kobresia humilis* meadow at slow-slope, 111 g C m^{-2} a^{-1} for the *Dasiphora fruticosa* shrub meadow and 73.3 g C m^{-2} a^{-1} for the *Arrhenatherum elatius*, respectively. The difference of the F_t value is notable among the investigated ecosystems. The F_t value of a single soil horizon decreases with depth in the study area (Figure 7). The ratio of the F_t value in the upper soil horizons (from 0 to10 cm) to the F_t value of the total profile is evidently different among the investigated ecosystems. For example, the CO_2 emissions from microbial decomposition of organic matter in the upper soil horizons (from 0 to10 cm) represents 81.23% of total F_t value for the *Kobresia humilis* meadow at bottomland, 72% for the *Kobresia humilis* meadow at slow-slope, 43.78% for the *Dasiphora fruticosa* shrub meadow and 40.38% for the *Arrhenatherum elatius*, suggesting that the CO_2 produced from microbial decomposition of organic matter in the upper soil horizons (from 0 to10 cm) is a major contributor to CO_2 efflux of total soil profile in the alpine meadow ecosystems, particularly for the *Kobresia humilis* meadow ecosystem. The soil-CO_2 flux for the *Dasiphora fruticosa* shrub meadow and the *Arrhenatherum elatius* gradually decline with depth, coinciding with the SOC distribution with depth.

The turnover rates of soil organic matter has observable difference among the studied ecosystems, attributing to different soil microclimate, life activity, physical-chemistry property of soil and human activity. The SOM turnover rate (from 0.0011 to 0.02216 a^{-1}) in the upper soil horizons (from 0 to 10 cm) for *Kobresia humilis* meadow is significantly higher than that (from 0.0021 to 0.0077 a^{-1}) in the *Dasiphora fruticosa* shrub meadow and the *Arrhenatherum elatius*. The decomposition rate of the SOM for the subsoil horizons (from 10 to 60 cm), however, is higher in the *Dasiphora fruticosa* shrub meadow and the *Arrhenatherum elatius* than that in the *Kobresia humilis* meadow. In addition, one factor resulting in the difference of SOC storage and its decomposition rate is the amount of aboveground litters and roots, and their chemical component among the different ecosystems in the study area, since the soil organic matter mostly originates from the aboveground litters and roots.

Up to date, the contribution derived from live root respiration and decomposition of soil organic matter to the total soil-CO_2 flux has not been confirmed. Most studies suggest that root respiration contributes some 30% to 70% of the total soil-CO_2 flux[43,45]. Root respiration contributes, on the average, 29% to the total soil CO_2 flux under forest ecosystem in Northeast China[41]. On a global scale, soil CO_2 flux correlates positively with mean annual temperatures and mean annual precipitation[43]. Mean annual air temperatures (−1.7℃) and mean annual precipitation (580 mm) in the study area are lower than that (2.7℃ and 600—800 mm, respectively) in the forest ecosystem of northeast China. Plants grow only from May to September, and non-grow season is up to seven months (from October to April), longer soil freezing time, and carbon dioxide emission from soil occurs negative from the first ten days of November month to the middle ten days of February month, i.e., the soil absorbs atmospheric CO_2 during the pe-

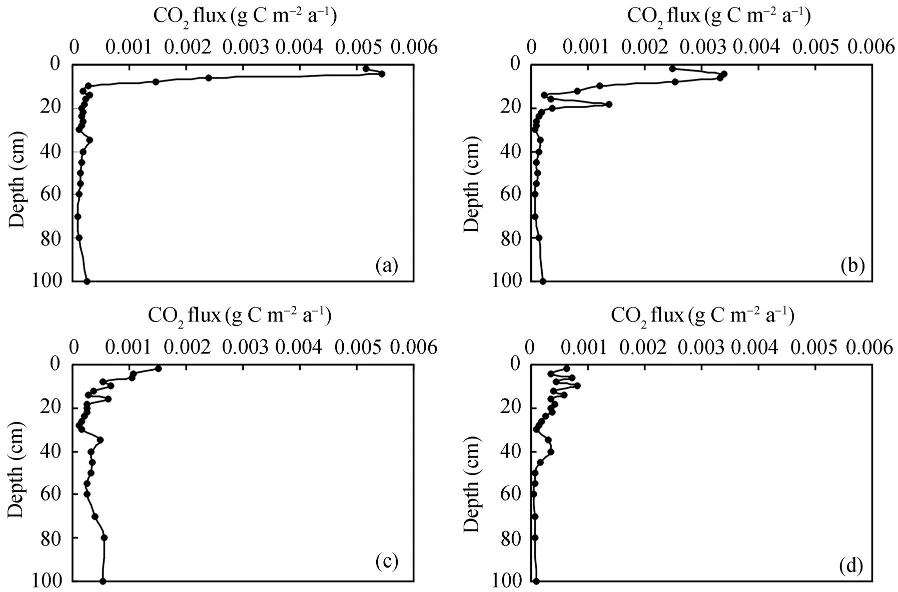

Figure 7 CO_2 production from the decomposition of soil organic matter vary with depth in different ecosystems. (a) ASC I profile; (b) ASC II profile; (c) JLM profile; (d) YMC profile.

riod[10]. Consequently, it is suggested that live root respiration contributes less 29% to the total soil CO_2 flux in this study area. If we assume that root respiration contributes 29% to the total soil CO_2 flux in this study area and, soil CO_2 is produced entirely from root respiration and decomposition of soil organic matter, then the soil-CO_2 flux was estimated at, in turn, 254.93 g C m^{-2} a^{-1} for the *Kobresia humilis* meadow at bottomland, 250.42 g C m^{-2} a^{-1} for the *Kobresia humilis* meadow at slow-slope, 156.34 g C m^{-2} a^{-1} for the *Dasiphora fruticosa* shrub meadow and 103.24 g C m^{-2} a^{-1} for the *Arrhenatherum elatius*, respectively, with an average of 191.23 g C m^{-2} a^{-1}. The soil CO_2 flux in the *Kobresia humilis* meadow is larger than that in the *Dasiphora fruticosa* shrub meadow, consistent with other studies[10,17], but this result is less than that from Cao (663 g C m^{-2} a^{-1} and 661 g C m^{-2} a^{-1} respectively)[10]. By comparison, the soil CO_2 flux in the Tibetan Plateau is significantly higher than that in Tundra (60 g C m^{-2} a^{-1}), and evidently less than that (457—488 g C m^{-2} a^{-1}) in adjacent latitude, low altitude North American great plain (38°50′N, 92°02′W)[43] and that (390—866 g C m^{-2} a^{-1}) in Inner Mongolia grassland[46]. This phenomenon results from the alpine environment on the Tibetan Plateau.

The organic carbon stored in the aboveground plants will increase, and roots and organic matter in the upper soil horizons will accumulate in the alpine meadow ecosystem due to the atmospheric CO_2 fertilization with global warming. However, it is found that the decomposition of organic matter was larger than the primary production in arctic tundra ecosystem by long-term nutrient fertilization, caused a net ecosystem loss of some 2000 g C m^{-2} over 20 years[47]. Hence, responding to global warming, the storage, volume of flow and fate of the soil organic carbon in the alpine meadow ecosystem of the Tibetan Plateau will be changed, which needs further research.

4 Conclusions

The alpine environment on the Tibetan Plateau plays a primary role for the accumulation of soil organic carbon. Soil organic carbon storage varies from 22.12×10^4 kg C hm^{-2} to 30.75×10^4 kg C hm^{-2}, with an average of 26.86×10^4 kg C hm^{-2} in the alpine meadow ecosystems. The soil carbon pool of the alpine meadow in the Tibetan Plateau plays an important role in Chinese or even in the global carbon budget.

Turnover times of organic carbon pools increase with depth from 45 a—73 a in the surface soil horizon to hundreds of years or millennia or even longer at the depth soil horizons in the alpine meadow ecosystems. Especially, more than 30% of total soil organic carbon resides in the relatively “active” pool at the topsoil horizons and is sensitive to global warming for the *Kobresia* meadow. Protection of soil coverage is a very important measurement for the *Kobresia* meadow ecosystem with the global climatic warming.

The size of soil organic carbon pool and its change are mostly controlled by soil environmental characteristics and vegetation types. The soil-CO_2 flux ranges from 103.24 g C m^{-2} a^{-1} to 254.93 g C m^{-2} a^{-1}, with an average of 191.23 g C m^{-2} a^{-1}. The CO_2 efflux from microbial decomposition of organic matter varies from 73.3 g C m^{-2} a^{-1} to 181 g C m^{-2} a^{-1}. 72.8%—81.23% of total CO_2 emitted from decomposition of organic matter results from the topsoil horizon (0—10 cm) for the *Kobresia* meadow. The CO_2 produced from decomposition of organic matter in the upper soil horizons (0—10 cm) is a major contributor to total CO_2 efflux in the alpine meadow ecosystem.

Responding to global warming, the storage, volume of flow and fate of the soil organic carbon in the alpine meadow ecosystem of the Tibetan Plateau will be changed, which needs further research.

The authors would like to thank extensive valuable comments provided by four anonymous reviewers and Dr. Wang Bixiang.

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