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REGULAR ARTICLE

# Patterns of SOC and soil $^{13}C$ and their relations to climatic factors and soil characteristics on the Qinghai–Tibetan Plateau

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**Abstract**
*Background and aims* SOC inventory and soil $\delta^{13}C$ were widely used to access the size of soil C pool and to indicate the dynamics of C input and output. The effects of climatic factors and soil physical characteristics and plant litter input on SOC inventory and soil $\delta^{13}C$ were analyzed to better understand the dynamics of carbon cycling across ecosystems on the Qinghai-Tibetan Plateau.
*Methods* Field investigation was carried out along the two transects with a total of 1,875 km in length and 200 km in width. Sixty-two soil profiles, distributed in forest, meadow, steppe, and cropland, were stratified sampled every 10 cm from 0 to 40 cm.
*Results* Our result showed that SOC density in forest and meadows were much higher than in steppe and highland barley. In contrast, $\delta^{13}C$ in forest and meadow were lower than in steppe and highland barley. Soil $\delta^{13}C$ tended to enrich with increasing soil depth but SOC decline. SOC and $\delta^{13}C$ (0–40 cm) were correlated with different climatic factors in different ecosystems, such that SOC correlated negatively with MAT in meadow and positively with MAP in steppe; $\delta^{13}C$ correlated positively with MAT in meadow and steppe; and $\delta^{13}C$ also tended to increase with increasing MAT in forest. Of the variation of SOC, 55.15 % was explained by MAP, pH and silt content and 4.63 % was explained by the interaction between MAT and pH across all the ecosystems except for the cropland. Meanwhile, SOC density explained 27.40 % of variation of soil $\delta^{13}C$.
*Conclusions* It is suggested that different climatic factors controlled the size of the soil C pool in different ecosystems on the Tibetan Plateau. SOC density is a key contributor to the variation of soil $\delta^{13}C$.

**Keywords** Soil organic carbon · Soil $\delta^{13}C$ · Climate factors · Soil characteristic · Qinghai-Tibetan Plateau

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## Introduction

Terrestrial ecosystems contain the largest organic carbon (C) and thus take most important role in balancing atmospheric $CO_2$ concentration (Briones et al. 2006). $CO_2$ released from the mineralization of soil organic matter (SOM) is regarded as a large part contribution

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to anthropogenic C-driven climate change (Sanderman et al. 2003). Therefore, the dynamic of SOM and the capacity of soil to accumulate and stabilize organic C in response to climate changes have drawn much attention (Rovira and Vallejo 2003; Llorente and Turrión 2010; Wick et al. 2009).

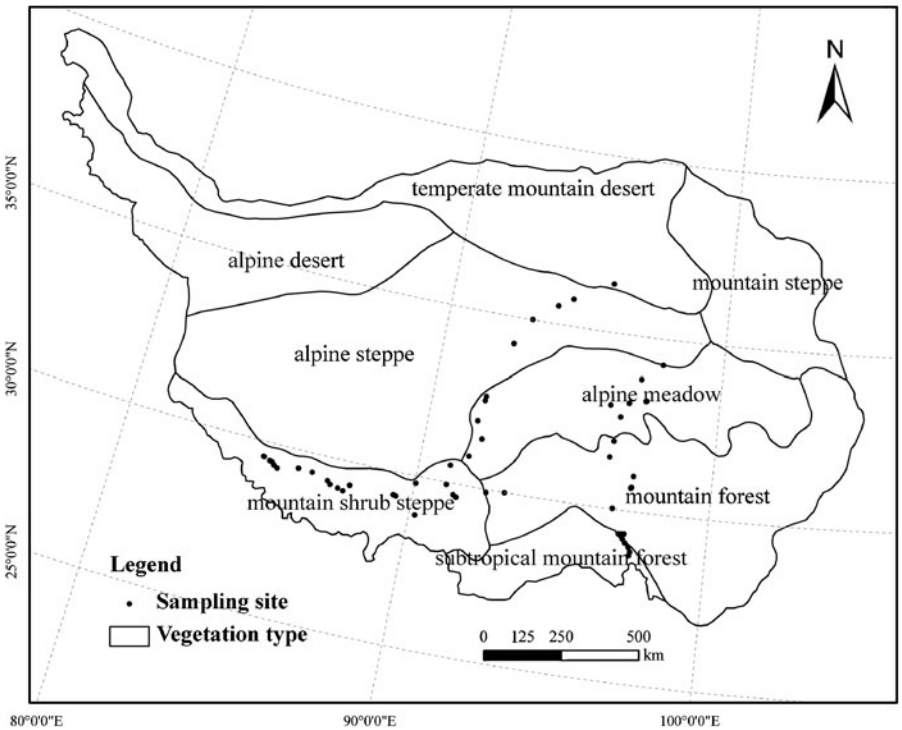
SOM is a complex mixture of material, and it exists in the soil profile as matters with different decomposition or stabilization states (Llorente et al. 2010). It has been suggested that differences in density fractionation of SOM represent different pools of SOM with different turnover rates (e.g. Swanston et al. 2002; Helfrich et al. 2006; Llorente and Turrión 2010). In general, active or labile fractions of SOM may be more temperature sensitive and respond much faster to climate changes than the older or recalcitrant fractions of SOM (Barrios et al. 1996).

Stable carbon isotope ($\delta^{13}C$) analysis has increasingly been used to estimate soil C turnover (e.g. Balesdent and Mariotti 1988; Bernoux et al. 1998; Glaser 2005; Sevink et al. 2005) and to assess the degree of SOM decomposition (e.g. Wedin et al. 1995; Connin et al. 2001). Stable isotope data are useful for assessing C turnover in SOM density fractions (Glaser 2005). Because the light fraction from density fractionation primarily represents labile material of relatively recent origin, $\delta^{13}C$ values of this soil fraction would reflect $\delta^{13}C$ values that are closer to current vegetation (Compton and Boone 2000; Accoe et al. 2002). Meanwhile, the heavy fraction, representing older, more recalcitrant and amorphous organic compounds, displays relatively enriched $\delta^{13}C$ values due to isotopic discrimination of the heavy isotope in SOM compounds during biological and/or physical processes (Andreux et al. 1990; Ehleringer et al. 2000).

Spatial patterns of the stable carbon isotopic composition ($\delta^{13}C$) of SOC have been reflected by a data set collected from different transects across the world, which included several natural gradients, such as climatic, edaphic, and biotic controls (Bird et al. 2001, 2002a, b, 2004; Wynn et al. 2006). For example, Bird et al. (2002b) presented SOC inventories and C isotope compositions in a 1000 km transect along the Yennisey River in central Siberia. Although previous studies provided a well general framework on the spatial patterns of soil stable C isotopic composition along environmental gradients at large scales, regional SOC and $\delta^{13}C$ data sets are currently incomplete. Because both soil characteristics and climatic conditions and vegetation are highly heterogeneous, better understanding the patterns of SOM and soil $\delta^{13}C$ and the controlling factors at regional scales are critical to accurately quantify the rate of C cycle and C storage in response to climate changes (Bird et al. 2001; Bird et al. 2002a, b, 2004; Lu et al. 2004).

The Qinghai-Tibetan Plateau (QTP) is called the "the third pole of the earth" and "world roof" owing to its high altitude (average altitude of more than 4000 m asl.). Climate conditions varied greatly across the area, with warm and humid climate in the southeast, and cold and arid climate in the northwest (Zheng 1996). Annual precipitation decreases gradually from the southeast to the northwest. Along these temperature and precipitation gradients, ecosystem types change gradually from marine humid montane (tropical seasonal and rain forest, warm-temperate broad leaved evergreen forest, temperate deciduous forest, and coniferous forest) in the southeastern to continental semiarid montane (temperate shrubland or meadow, temperate steppe, alpine meadow or shrubland, and alpine steppe) in the middle region to continental arid montane (temperate desert, alpine desert, and ice or polar desert) in the northwestern region (Ni 2000) (Fig. 1). The QTP accumulated large amounts of SOC, therefore, might serve as a sink of atmospheric $CO_2$ (Ni 2002; Kato et al. 2004). However, with global changes, the plateau is experiencing climatic warming (Wang et al. 2001, 2003; Wu et al. 2005) and the increasing rate in mean average temperature (MAT) has been estimated amount to 0.45–0.70 °C per 10 yrs between 1960 and 2000 (Wang et al. 2001). This implies that mineralization of SOM might be enhanced and give rise to increase in $CO_2$ emission from the soil of the plateau. Recently, more and more works have been carried out on the plateau to investigate SOM dynamic and soil capacity to accumulate and stabilize organic C (Wang et al. 2003; Yang et al. 2008). To date, most of these studies were conducted in a very restricted range, and very few transect studies were performed to examine patterns of both SOC and soil $\delta^{13}C$ in regional scale synchronously. The lack of knowledge on patterns of SOC and soil $^{13}C$ both in space and in depth makes it difficult to understand the response of soil C cycles to climate warming across the plateau. Meanwhile, plant tissue is the main source of SOM, and plant $^{13}C$ should be an important factor influencing soil $\delta^{13}C$. Recent studies showed that

Fig. 1 Patterns of typical ecosystems on the Tibetan Plateau and sampling sites




changes in plant-derived C caused chemical alteration of non-aggregated particulate soil organic matter and influenced stability of SOM following grassland-to-woodland transition in a subtropical savanna (Filley et al. 2008; Crow et al. 2009).

Here, we present a data set of SOC and soil $^{13}C$ and leaf $^{13}C$ composition across ecosystems on the Qinghai-Tibetan Plateau. Ecosystem types, such as, steppes (desert steppe, temperate steppe, alpine steppe), meadows (alpine meadow, mountain meadow, lowland meadow), cropland (highland barley) and forest, were concerned. We try to (1) seek the patterns of SOC and soil $\delta^{13}C$ natural abundance among ecosystem types, (2) explore the vertical patterns of SOC and soil $^{13}C$ along with soil depth, and (3) to examine the effects of climate factors and soil physical characteristics and leaf $\delta^{13}C$ of dominant species on the patterns of SOC and soil $\delta^{13}C$. Integrated above results we try to test the hypothesis that the patterns of soil $\delta^{13}C$ along soil profiles indicate degree of SOM decomposition, and the spatial patterns of soil $\delta^{13}C$ across all the ecosystems indicate dynamics of C input and output which strongly controlled by the gradient of precipitation or temperature from northwest to southeast of the plateau.

## Materials and methods

### Ecological transects

Two transects, a total of 1,875 km in length and 200 km in width, were set up on the Qinghai-Tibetan Plateau (latitudes from 28.73 to 36.38 º N, longitudes from 84.83 to 97.85 º E, and altitudes from 2350 to 5500 m). 62 soil profiles were sampled which covered by desert steppe, temperate steppe, alpine steppe, alpine meadow, montane meadow, lowland meadow, highland barley and forest from the northeast to the southwest. The locations of the sampling sites are showed in Fig. 1. The transects extended from the Bayan Har Mountains in Qinghai Province (34.26 ° N, 97.85 ° E) on eastern end to Saga County in Tibet Province (29.57 ° N, 84.83 ° E) on western end. The average altitude is about 4143 m asl. across the transects (Fig. 1). MAT increases from −5.3 to 12.6 °C and MAP increases from 204 to 995.1 mm from north to south along the transects. There is no large temperature gradient from west to east (84–99 ºE) and variation in mean annual temperature (MAT) is about 2 °C. On the contrary, the overall annual precipitation (MAP) increases from 300 to 600 mm along west to east. The zonal soils are mostly equivalent to the Cryic Aridisols in terms of the FAO-UNESCO system. The soils are characterized with coarse texture and loose structure, which are highly susceptible to wind erosion (Zou et al. 2002). Grassland is the main land cover patterns in the study area, amount to 76.2 % of the total land area (Fan et al. 2008).

Field sampling and data collection

The 62 sites were selected in areas that had no timber harvest or grazing, and thus try to reduce anthropogenic disturbance except cropland. Highland barley has been cultivated for more than 20 years after transformation of local vegetation montane steppe into the cropland. Systematic sampling method was performed with interval between sites at about 50 km. But some sites were more widely spaced and the average distance exceeded 50 km owing to the tough topography constraint. Furthermore, we adjusted sampling sites to include grassland types omitted by the systematic sampling. Geographic location of each site was determined by a Global Positioning Satellite device. Meanwhile, altitude, soil condition, and vegetation type were recorded.

All soil samples were collected from July to September in 2004 and 2005. For most sites three $1 \times 1$ $m^2$ quadrats were set up for plant community survey. For alpine meadow, three $0.25 \times 0.25$ $m^2$ quadrats were used (Fan et al. 2008). Leaf samples in grassland were collected and used for measurement of leaf $\delta^{13}C$ (to see Zhou et al. 2011 for the detailed information). In each quadrat, two to three sites were randomly selected for soil sampling, and soils were stratified sampled every 10 cm in depth from 0 to 40 cm in each site. The soil corer was pounded into the soil to the specified depth (10, 20, 30, and 40 cm, respectively) after removal of the surface litter. 9–36 soil cores in 3.1 cm diameter were collected in each site, but some samples were obtained by digging out $25 \times 25$ $cm^2$ pits. A total of 268 samples were collected across the two transects. Here we only concerned the upper 40 cm in depth of the soil. Samples were stored in bags and transported to the laboratory for further analysis.

Soil samples were air-dried and then sieved through 2 mm screen to remove root, stone, and other debris. Soil particles were defined based on the percentage of clay (< 0.002 mm), silt (0.002–0.05 mm) and sand (0.05–2.0 mm), respectively. Sand fractions were measured using wet sieving. Silt and clay fractions were measured using the pipette method after removing organic matter. Samples were digested in a heated hydrogen peroxide solution, and then dispersed using sodium hexamethaphosphate (ISSCAS 1978; Su et al. 2002). Part of each air-dried sample was finely ground to pass through 1 mm sieve for chemical and isotopic analysis.

Soil samples were acidified with 0.1 M HCl to remove carbonates. C concentrations and stable C isotope ratios were measured using elemental analysis-continuous flow mass spectrometry with PDB (for Pee Dee Belemnite) as reference (Finnegan MAT253, Thermal Electron Corporation, Waltham, Massachusetts, USA). $^{13}C$ isotope analysis was performed in the Institute of Geographic Sciences and Natural Resources Research. The overall analytical precision is less than 0.1‰.

Most of leaf $\delta^{13}C$ values (16 values in meadow, 5 values in alpine steppe, and 10 values in temperate steppe) were provided by Zhou (Zhou et al. 2011) who was responsible for plant sampling in grassland in the field investigation. Leaf $\delta^{13}C$ values in desert steppe (5 values) and forest (5 values) were provided by Song (Song et al. 2008) according to geographic location and vegetation types. No data of leaf $\delta^{13}C$ were obtained in cropland. MAT and MAP values for all sampling sites were extracted from the climate database (Yu et al. 2004), which interpreted based on national meteorological observation data from1960 to 2000.

Data analysis

Of the 65 soil profiles, one missed $\delta^{13}C$ values in the surface (alpine steppe), another two missed $\delta^{13}C$ values in the middle layer (lowland meadow). Thus, SOC and $\delta^{13}C$ values in 62 soil profiles were used in this study.

We use one-way ANOVA to analyze the differences in SOC and soil $\delta^{13}C$ among ecosystems as well as among soil layers. Differences in leaf $\delta^{13}C$ among ecosystems were analyzed using one-way ANOVA. Curve estimation of regression analysis was used to test the relationship of SOC and soil $\delta^{13}C$ with climatic factors (MAT, MAP) across and within all the four ecosystems, respectively. We also used curve estimation of regression analysis to examine the relationships of SOC and soil $\delta^{13}C$ with soil physical characteristics (pH and sand, silt, clay content). In addition, the relationship between soil $\delta^{13}C$ and SOM was examined using regression analysis. We used GLM analyses to examine the correlation of soil δ$^{13}$C and SOC with climatic factors and leaf δ$^{13}$C and soil characteristic (pH, and sand, slit, and clay content) and their interactions across all the ecosystems except for cropland. All the analyses were performed using SPSS version 14.0 (SPSS Inc. Chicago).

## Results

### Patterns of SOC, soil δ$^{13}$C and leaf δ$^{13}$C in different ecosystems

Forest soil (0–40 cm) had an average SOC density of 4.61 kg C m$^{-2}$ which is comparable with the SOC density of 4.52 kg C m$^{-2}$ in meadows (Table 1). SOC density did not differ between steppe (1.64 kg C m$^{-2}$) and cropland (2.02 kg C m$^{-2}$), which were much lower than that in forest and meadow soils (Table 1). On the contrary, soil δ$^{13}$C values in steppe and cropland were higher than in forest and meadows (Table 1). There were no differences in δ$^{13}$C between forest (−24.66‰) and meadows (−24.55‰) as well as between steppe (−22.88‰) and cropland (−23.44‰) (Table 1). Leaf δ$^{13}$C of dominant species was lower in forest than in meadows and steppes ($F_{2,39}$=13.12, $P$<0.0001; Table 1).

### Vertical patterns of SOC and soil δ$^{13}$C in different ecosystems

SOC density showed decreasing trends with increase in soil depth from 0 to 40 cm in all the ecosystems (Table 2). In each soil layer, SOC density was much higher in forest and meadows than that in steppes and cropland (Table 2). In contrast to SOC density, soil δ$^{13}$C showed increasing trends with soil depth from 0 to 40 cm in all the ecosystems (Table 2), although the increasing trends were statistically significant only in meadow ($F_{3,26}$=8.159, $P$<0.013) and highland barley ($F_{3,39}$=10.070, $P$=0.0001). In each layer, soil δ$^{13}$C were higher in steppe than in other ecosystems ($F_{7,65}$=13.843, $P$<0.0001) (Table 2).

### Relationship of SOC and soil δ$^{13}$C with climatic factors and soil characteristics and leaf δ$^{13}$C

The relationship of SOC density and soil δ$^{13}$C (0–40 cm) with MAT and MAP were not significant across all the ecosystems. We found that SOC density and soil δ$^{13}$C were correlated with different factors in different ecosystems. In meadow, SOC density was negatively correlated with MAT, but soil δ$^{13}$C was positively correlated with MAT, and both SOC density and δ$^{13}$C was not correlated with MAP (Fig. 2a–d). SOC density was positively correlated with MAP, but not with MAT in steppe (Fig. 2e, f). Contrarily, soil δ$^{13}$C was positively correlated with MAT, but not with MAP in steppes (Fig. 2g, h). In forest, relationship of SOC density and δ$^{13}$C with MAT and MAP was not significant (Fig. 2i–l).

Relationship of SOC density and soil δ$^{13}$C with pH and soil texture was not significant within each of the four ecosystems. But we found some general relationships of SOC density and soil δ$^{13}$C with pH and soil texture across all the ecosystems. For example, SOC density (0–40 cm) exponentially negatively correlated with pH, but we did not find significant relationship between soil δ$^{13}$C (0–40 cm) and pH across all the ecosystems (Fig. 3a, e). SOC density was exponentially negatively correlated with sand content, but exponentially positively with silt content (Fig. 3b, c). On the contrary, soil δ$^{13}$C was positively correlated with sand content, and negatively with silt content (Fig. 3f, g). The relationship of δ$^{13}$C with clay content was not statistically significant (Fig. 3h). Soil δ$^{13}$C (0–40 cm) was negatively correlated with logarithmically transformed SOC (0–40 cm) across all the ecosystems (Fig. 4a), and

**Table 1** SOC density and soil δ$^{13}$C values in 0–40 cm depth among ecosystem types on the Qinghai-Tibetan Plateau. Values are mean ±s.e. n are the number of soil profiles. Values sharing the same letters are not different at $P$=0.05 (by Duncan test)

| Ecosystem types | n | SOC (kgC/m$^2$) | Soil δ$^{13}$C (‰) | Leaf δ$^{13}$C (‰) |
|---|---|---|---|---|
| Forest | 10 | 4.61±0.65b | −24.66±0.26a | −27.78±0.40a |
| Meadow | 18 | 4.52±0.50b | −24.55±0.14a | −26.61±0.13b |
| Steppe | 24 | 1.64±0.24a | −22.88±0.39b | −26.05±0.20b |
| Cropland | 10 | 2.02±0.29a | −23.44±0.41b | no data |

**Table 2** SOC density and soil $\delta^{13}C$ values in each layer (10 cm interval) from 0 to 40 cm depth in different ecosystem types on the Qinghai-Tibetan Plateau. Values are mean±s.e. Values sharing the same letters are not different at $P$=0.05 (by Duncan test)

| Depth | Forest ($n$=10) | Meadow ($n$=18) | Steppe ($n$=24) | Cropland ($n$=10) |
|---|---|---|---|---|
| SOC (kgC/m$^2$) | | | | |
| 0–10 cm | 7.38±0.91b | 6.11±0.65b | 2.33±0.44b | 2.70±0.35c |
| 10–20 cm | 4.78±0.82a | 4.84±0.56ab | 1.86±0.29ab | 2.39±0.34bc |
| 20–30 cm | 3.47±0.57a | 3.77±0.52a | 1.20±0.16a | 1.65±0.32ab |
| 30–40 cm | 2.80±0.67a | 3.34±0.47a | 1.14±0.17a | 1.34±0.21a |
| $\delta^{13}C$ (‰) | | | | |
| 0–10 cm | −25.25±0.33a | −24.87±0.19a | −23.65±0.30a | −24.35±0.39a |
| 10–20 cm | −24.50±0.30a | −24.71±0.18a | −22.79±0.46a | −23.98±0.43a |
| 20–30 cm | −24.48±0.29a | −24.54±0.15ab | −22.56±0.45a | −23.12±0.49ab |
| 30–40 cm | −24.40±0.25a | −24.08±0.18b | −22.52±0.43a | −22.33±0.47b |

the same trends were found in each layer of the soils (Fig. 4b–e).

GLM analysis suggested that SOC density, MAT, altitude, and leaf $\delta^{13}C$ account for 45.21 % of the overall variation of soil $\delta^{13}C$ in the top 0–40 cm, and interaction between SOC density and altitude accounted for 7.76 % of the variation in soil $\delta^{13}C$ across all the ecosystems except for cropland (Table 3). Meanwhile, MAP, pH and silt content explained 55.15 % of the overall variation of SOC density in the top 0–40 cm, and interaction between MAT and pH explained 4.63 % of the variation of SOC density (Table 3).

## Discussion

We found general increase in soil $\delta^{13}C$ with depth in each ecosystem, and significant correlations of soil $\delta^{13}C$ with SOC density, MAT, altitude, and leaf $\delta^{13}C$ across the ecosystems. Our results support the hypothesis that the patterns of soil $\delta^{13}C$ along soil profiles indicate degree of SOM decomposition, and the spatial patterns of soil $\delta^{13}C$ across all the ecosystems indicate dynamics of C input and output which strongly controlled by the gradient of precipitation or temperature from northwest to southeast of the plateau. However, we found that different climatic factors controlled the accumulation of SOC and the degree of soil organic matter decomposition in different ecosystems. In alpine meadow and alpine steppe, the significant positive correlations of soil $\delta^{13}C$ with temperature (Fig. 2c, g) suggest that low temperature is the key factor limiting the degree of SOM decomposition, although SOC accumulation in alpine meadow was mainly limited by temperature and in alpine steppe by precipitation, respectively (Fig. 2a, f).

### SOC and soil $\delta^{13}C$ in different ecosystems

Forest and meadow soils contain much more SOC than steppe and cropland soils on the Qinghai-Tibetan Plateau. Yang et al. (2010) found that SOC in the top 20 cm accounted for about 55 % and 41 % of the total SOC in the 1 m depth in alpine meadow and alpine steppe, respectively, on the plateau. In our study, SOC in the upper 20 cm accounted for about 50 % of the total SOC in 40 cm depth of soil among all ecosystems. SOC density in meadow (4.52 kg C m$^{-2}$) and steppe (1.64 kg C m$^{-2}$) in our study was comparable with those simulated by CENTURY model which showed that SOC density was 3.87 and 1.32 kg C m$^{-2}$ in alpine meadow and alpine steppe, respectively (Zhang et al. 2007). Our results also indicate that SOC density in meadows was 2–3 times higher than that of steppes on the plateau. The potential reason may be that the annual precipitation in meadows (479 mm) was higher than that of steppes (372 mm), which led to higher biomass in meadows. In addition, lower mean annual temperature in meadows (0.6 °C) relative to steppes (1.1 °C) constrained decomposition of SOM and contributed accumulation of SOC in meadow soils, which is also confirmed by the lower $\delta^{13}C$ values in meadows. In addition, clay (silt) to sand ratio is much higher in the meadow than

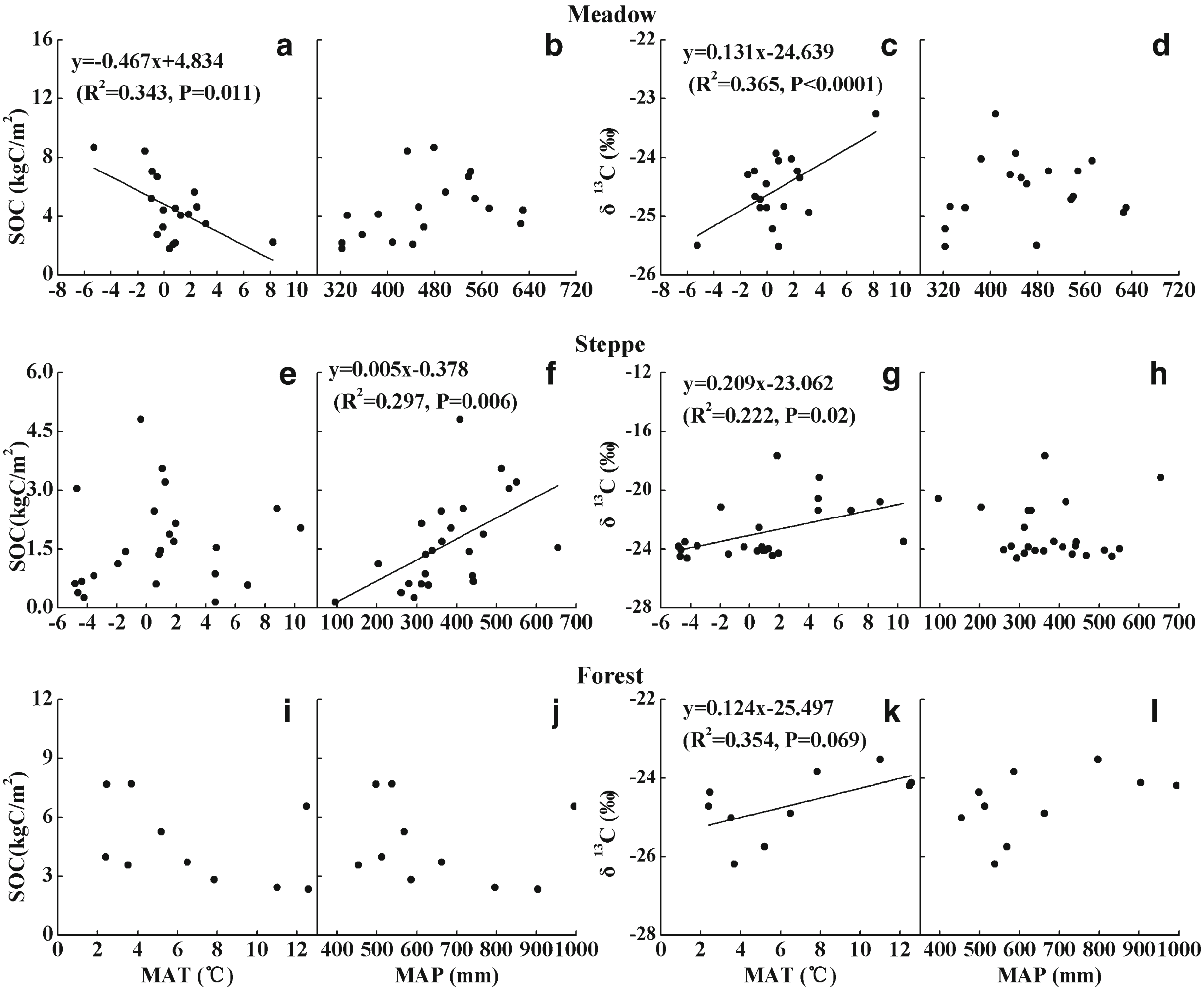



**Fig. 2** Relationships of SOC density and soil $\delta^{13}C$ with mean annual temperature and annual precipitation in meadows (**a**–**d**), steppes (**e**–**h**), and forests (**i**–**l**)

steppe which would promote micro-aggregation and clay-SOM interactions thus slowing decay.

Lu et al. (2004) found that the soil $\delta^{13}C$ values (0–5 cm) ranged from −26.8 to −15.8‰ and exhibited a clear relation with ecosystem types, i.e. forest (−24.49 ‰), shrub (−24.7 ‰), steppe (−23.1 ‰), alpine meadow (−23.6 ‰), alpine desert steppe (−21.3 ‰), and alpine desert (−18.9 ‰). In our study, $\delta^{13}C$ values at the 0–10 cm soil interval ranged from forest of −24.66‰, alpine steppe of −22.88‰, Alpine meadow of −24.55‰, which is similar to the values measured by Lu et al. (2004). In addition, a clear increasing trends of soil $\delta^{13}C$ were detect from forest (−24.66 ‰), meadow (−24.55 ‰) to steppe (−22.88 ‰) which corresponding to leaf $\delta^{13}C$ values of dominant species in forest (−27.78 ‰), meadow (−26.61 ‰) and steppe (−26.05‰). Leaf $\delta^{13}C$ of dominant species accounted for 6.68 % of the variation in soil $\delta^{13}C$ across all the ecosystems except for cropland on the QTP. Leaf litter is the main source of SOM, and thus different plants with large differences in leaf $\delta^{13}C$ influence soil $\delta^{13}C$. Plant species growing in dry habitat generally have higher leaf $\delta^{13}C$ value and higher water use efficiency (Francey and Farquhar 1982; Silim et al. 2001; Warren et al. 2001). Leaf $\delta^{13}C$ of $C_4$ plants is much higher than $C_3$ plants, and $C_4$ plants have strong ability in adaptation to severe drought. Isotopic measurement on herbaceous species showed that $C_4$ plants exist in the high elevation QTP, although mostly in low

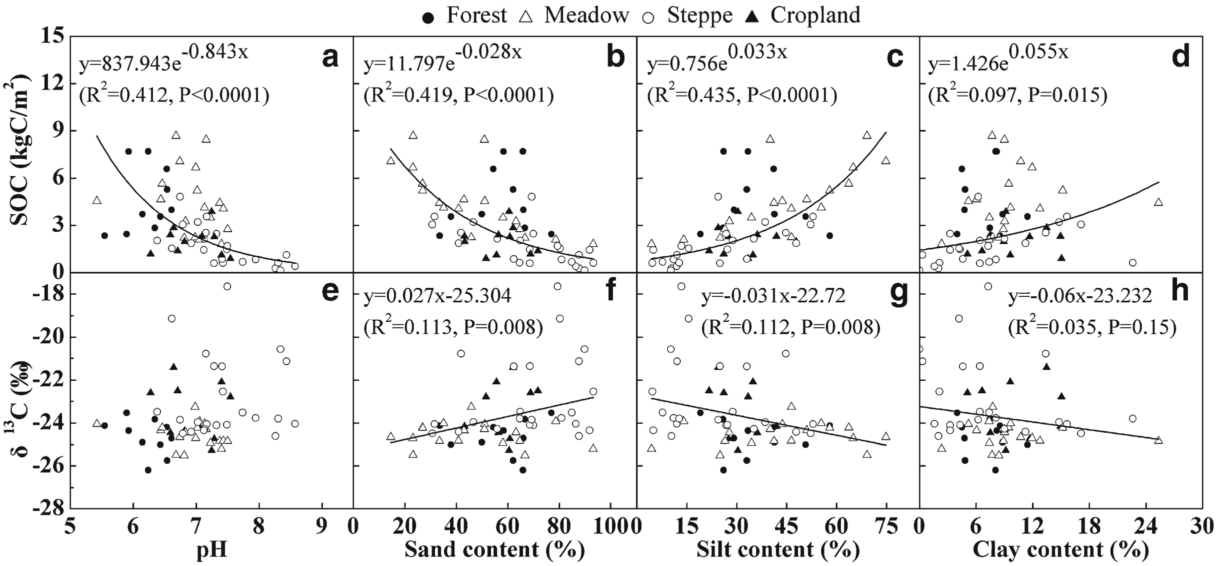



**Fig. 3** Relationships of SOC density (**a**–**d**) and soil $\delta^{13}C$ (**e**–**h**) with soil physical characteristics (pH, sand, silt, and clay content) across the four ecosystem types

abundances (Wang et al. 2004). However, in our study the highest leaf $\delta^{13}C$ value is −24.21‰ occurred in desert steppe. Our samples did not include $C_4$ plants.

## Vertical distribution of SOC and soil $\delta^{13}C$

A general increasing tendency of $\delta^{13}C$ with soil depth has been widely observed (Nadelhoffer and Fry 1988; Buchmann et al. 1997; Ehleringer et al. 2000; Feng 2002; Peterson et al. 2006). In General, $\delta^{13}C$ values in the uppermost soil layers directly reflect inputs of C from local standing biomass, but systematic changes with soil depth reflect more complex processes of microbial degradation and the mixing of SOC of different ages (Bird et al. 2004). The increase in $\delta^{13}C$ is due largely to the increasing residence time of the organic C in the soil (Bird et al. 2002b**)**. Bird et al. (2004) reported that $\delta^{13}C$ values increased linearly at both the 0–5 and 0–30 cm depth intervals in Botswana. Other studies found that approximately 1–2 ‰ variations in SOC $^{13}C$ generally occurred with soil depth (Nadelhoffer and Fry 1988; Wedin et al. 1995). Moreover, Stevenson et al. (2005) also showed 1–2 ‰ enrichment in SOC $^{13}C$ with depth to 1 m. Our finding that soil $^{13}C$ in deeper layers is more enriched on the plateau, and all the sites showed 1–1.2 ‰ enrichment in mean SOC $^{13}C$ with depth to 40 cm, which is consistent with previous studies. Specifically, the enrichment of soil $\delta^{13}C$ values with depth to 40 cm in forest, meadow, steppe and highland barley were 0.85, 0.79, 1.13 and 2.02 respectively. The highest enrichment in $\delta^{13}C$ value occurred in cropland (highland barley). We presumed that the transform of vegetative cover from original montane desert steppe into barley contributed to the variation in $\delta^{13}C$. Unfortunately, we did not measure leaf $\delta^{13}C$ of barley, additionally, knowledge about process of SOM decomposition is shortage. So the legacy effect of the original vegetation on soil $\delta^{13}C$ need to be further investigated.

## Effects of climate factors on SOC and soil $\delta^{13}C$

The plateau has complex physical geography and climate. Variations in precipitation and temperature are crucial factors controlling the patterns of vegetation on the plateau (Lu et al. 2004). Although large variation in leaf $\delta^{13}C$ of dominant species reflected ecosystem patters along precipitation gradient (Song et al. 2008), spatial patterns of soil $\delta^{13}C$ did not show strong correlations with MAT or MAP across ecosystem types. Although soil $\delta^{13}C$ did not show consistent variation with precipitation or temperature gradients from northwest to southeast of the plateau in regional scale, different climatic factors controlled soil SOC and $\delta^{13}C$ in different ecosystems. The significant negative correlation of SOC with MAT and positive correlation of soil $\delta^{13}C$ with MAT in meadows indicated that MAT is key factor influencing

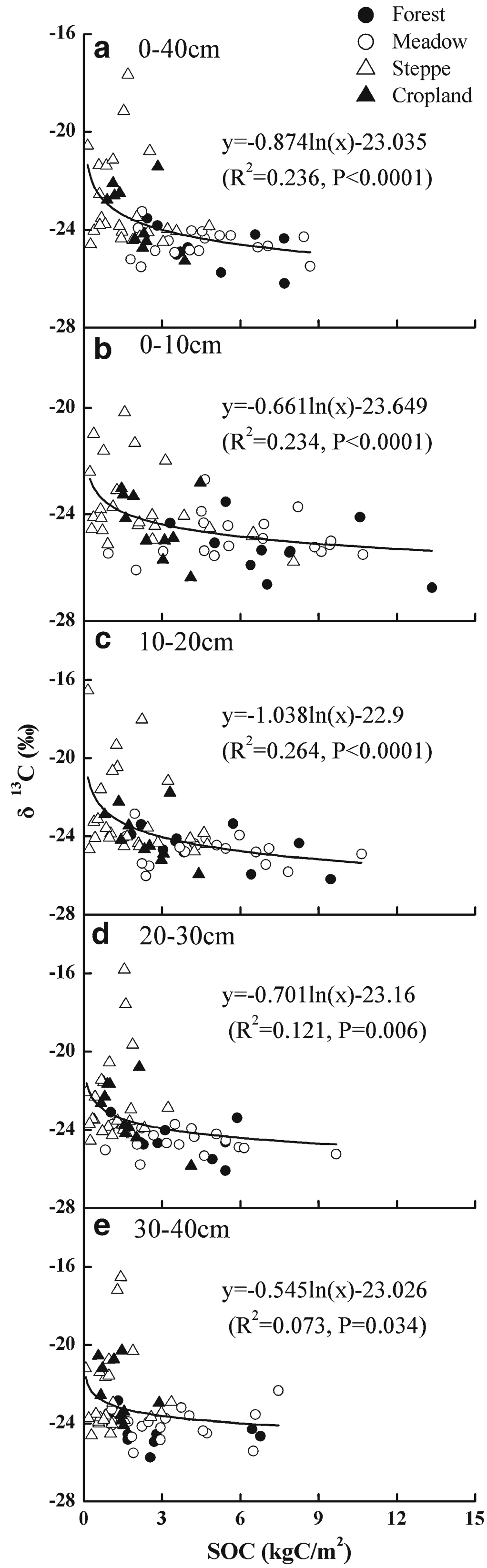



**Fig. 4** Relationships of soil $\delta^{13}C$ with SOC density in (**a**) 0–40 cm, (**b**) 0–10 cm, (**c**) 10–20 cm, (**d**) 20–30 cm, and (**e**) 30–40 cm soil layers across the four ecosystem types, respectively

SOC inventory and its decomposition and turnover in meadow ecosystem on the plateau. Meanwhile the significant positive correlation of SOC with MAP and positive correlation of $\delta^{13}C$ with MAT in steppe indicate that precipitation and temperature are key factors controlling SOC accumulation and SOC decomposition and turnover, respectively, in steppes. Bird et al. (2002b) reported that temperature and precipitation strongly affected distribution of organic C in the soil and its isotopes along a north-south transect in western Canada. On the QTP, temperature was the critical climate factor in controlling $CO_2$ exchange. Low temperature constrained the decomposition of litter and limited ecosystem respiration (Kato et al. 2004). Furthermore, Shi et al. (2006) also demonstrated that soil temperature is an important factor controlling temporal variation of soil $CO_2$ efflux at an observation site on the plateau. Meanwhile, precipitation is another important factor controlling SOC inventory. In meadows, low temperature and high precipitation contributed to high SOC content and slow decomposition and turnover of organic matter (Kato et al. 2004). In steppes, precipitation is a key factor limiting plant growth. It has been suggested that soil moisture take an important role in shaping spatial patterns of SOC density in alpine grasslands (Yang et al. 2008). In our study, 25.86 % of the variation of SOC was explained by MAP across all the ecosystems, indicated soil moisture influences soil SOC across both grassland and forest in QTP. We also found that temperature and the interaction between temperature and pH contributed to variation of soil $\delta^{13}C$ and SOC, respectively, indicated that temperature influences dynamics of C input and output in ecosystems on the QTP.

## Effect of soil texture on SOC and soil $\delta^{13}C$

Soil texture is an important index reflecting numerous soil properties. Generally, organic matter breaks down faster in sandy soils than in fine-textured soils, given similar environmental conditions, tillage, and fertilizer inputs (Ladd et al. 1993; Schjønning et al. 1999; Thomsen et al. 1999). Wynn et al. (2006) reported that soil texture influences SOC inventory. Moreover, Yang et al. (2008) indicated that SOC in alpine grasslands significantly increased with clay and silt content.

**Table 3** General linear model (GLM) analysis for the integrative effects of climatic factors (mean annual temperature, annual precipitation), altitude, leaf $\delta^{13}C$ and soil characteristics (pH, and sand, silt and clay content) on soil $\delta^{13}C$ and SOC in the top 40 cm

| | Source | df | Parameters | SE | MS | SS(%) |
|---|---|---|---|---|---|---|
| Soil $\delta^{13}C_s$ | SOC | 1 | −10.96 | 3.11 | 24.65*** | 27.40 |
| | MAT | 1 | 0.036 | 0.013 | 3.88* | 4.31 |
| | Alti | 1 | −0.020 | 0.011 | 6.14** | 6.82 |
| | $\delta^{13}C_l$ | 1 | 3.52 | 1.83 | 6.01** | 6.68 |
| | SOC × Alti | 1 | $1.86\times10^{-3}$ | $0.99\times10^{-3}$ | 6.98** | 7.76 |
| | Residuals | 30 | | | 1.15 | 38.50 |
| SOC | MAP | 1 | 0.005 | 0.001 | 58.19*** | 25.86 |
| | pH | 1 | −5.35 | 1.41 | 40.96*** | 18.20 |
| | Silt | 1 | 0.21 | 0.10 | 24.96** | 11.09 |
| | MAT × pH | 1 | −0.27 | 0.11 | 10.42* | 4.63 |

$\delta^{13}C_s$ represented for soil $\delta^{13}C$, SOC for soil organic carbon density, MAT for mean annual temperature, MAP for annual precipitation, Alti for Altitude, $\delta^{13}C_l$ for leaf $\delta^{13}C$, pH for soil pH value, Silt for soil silt content, df for degree of freedom, SE for standard errors, MS for mean squares, SS for proportion of variances explained by the variable. The parameters of the best-fit GLM model and their standard errors (SE) were presented. * $P<0.05$, ** $P<0.01$, *** $P<0.001$.

Although SOC content (0–40 cm) did not significantly increase with soil clay content, our results demonstrated that SOC storage (0–40 cm) positively correlated with silt content and negatively correlated with sand content across all the ecosystems. This confirms that fine-textured soil contained higher SOC in the plateau.

Soil texture influenced soil $\delta^{13}C$ values (Bird et al. 2002a). Generally, soil $\delta^{13}C$ increases with decreasing particle size, because C enriched in $\delta^{13}C$ is partitioned into microbial biomass and can subsequently be stabilized by interactions with fine mineral phases in the soil profile (Bird et al. 2002b; Sollins et al. 2009; Kleber et al. 2010). Bird et al. (2002b) also reported that organic C in the 0–5 cm layer of soils transfers from standing biomass into the coarsest size fractions of the soil. It is then degraded over time, with the residue progressively transferred into the more-resistant finer particles. The significant negative correlation of $\delta^{13}C$ with silt content and positive correlation of $\delta^{13}C$ with sand content in our study don't agree with those of Bird et al. (2002b). The reason may be that soils on the QTP developed slowly owing to low temperature. Soil layer with abundant organic matter is shallow and silt content decrease with increasing depth. But old soils in deeper layers tend to be enriched in $^{13}C$ (Liao et al. 2006a, b). Therefore, soil $\delta^{13}C$ value is positively correlated with sand content in our study transect.

In general, decomposition processes and turnover rate of C are the primary factors affecting SOC $^{13}C$ (Bird et al. 2001, 2002a, b, 2004). Some studies have found that a linear relationship exists between SOC inventories and $\delta^{13}C$ values (0–5 cm interval) along transects (Bird et al. 2001). Similarly, a significant negative relationship between $\delta^{13}C$ values and SOC inventories existed in soil profile form 0 to 40 cm, and also existed in each soil layer across all the ecosystems on the QTP.

In summary, this study demonstrates that different climatic factors controlled the accumulation of SOC and the degree of SOM decomposition in different ecosystems on the Qinghai-Tibetan Plateau. However, we did not find significant relationship of SOC and $\delta^{13}C$ with soil characteristics within each of the ecosystems. But SOC and $\delta^{13}C$ were correlated with soil characteristics across all the ecosystems. Given the positive relationship between temperature and soil $\delta^{13}C$ in alpine meadow and alpine steppe, and the increasing tendency of soil $\delta^{13}C$ with increasing temperature in forest, changes in decomposition and turnover rate of soil organic matter in different ecosystems should be pay more attention under global warming background. Although our soil samples were taken from different regions and ecosystem types on the Qinghai-Tibetan Plateau, the spatial representation of the sampling sites wasn't sufficient to cover spatial variability of SOC and $\delta^{13}C$ values across the wide area of the plateau. The regional relationship between $\delta^{13}C$ values and soil texture obtained from our study differs from that of previous studies and global patterns**.** Hence, mechanism of soil $^{13}C$ enrichment with soil depth needs to be explored. Moreover, in future studies, the effects of soil texture on soil $\delta^{13}C$ and age of soil substrate on soil $^{13}C$ abundance should be assessed separately.

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