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Temperature exerts no influence on organic matter δ^{13}C of surface soil along the 400 mm isopleth of mean annual precipitation in China

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Abstract. Soil organic carbon is the largest pool of carbon in the terrestrial ecosystem, and its isotopic composition is affected by a number of factors. However, the influence of environmental factors, especially temperature, on soil organic carbon isotope values ($\delta^{13}C_{SOM}$) is poorly constrained. This impedes the application of the variability of organic carbon isotopes to reconstructions of paleoclimate, paleoecology, and global carbon cycling. Given the considerable temperature gradient along the 400 mm isohyet (isopleth of mean annual precipitation – MAP) in China, this isohyet provides ideal experimental sites for studying the influence of temperature on soil organic carbon isotopes. In this study, the effect of temperature on surface soil δ^{13}C was assessed by a comprehensive investigation of 27 sites across a temperature gradient along the isohyet. Results demonstrate that temperature does not play a role in soil δ^{13}C. This suggests that organic carbon isotopes in sediments cannot be used for paleotemperature reconstruction and that the effect of temperature on organic carbon isotopes can be neglected in the reconstruction of paleoclimate and paleovegetation. Multiple regressions with MAT (mean annual temperature), MAP, altitude, latitude, and longitude as independent variables and $\delta^{13}C_{SOM}$ as the dependent variable show that these five environmental factors together account for only 9 % of soil δ^{13}C variance. However, one-way ANOVA analyses suggest that soil type and vegetation type are significant factors influencing soil δ^{13}C. Multiple regressions, in which the five aforementioned environmental factors were taken as quantitative variables, and vegetation type, soil type based on the Chinese Soil Taxonomy, and World Reference Base (WRB) soil type were separately used as dummy variables, show that 36.2, 37.4, and 29.7 %, respectively, of the variability in soil δ^{13}C are explained. Compared to the multiple regressions in which only quantitative environmental variables were introduced, the multiple regressions in which soil and vegetation were also introduced explain more of the isotopic variance, suggesting that soil type and vegetation type exert a significant influence on $\delta^{13}C_{SOM}$.

1 Introduction

While global climate change has received a great deal of attention in recent years, effective predictions of future climate change depend on relevant information about climate in the geological past. Over recent decades, stable carbon isotopes in sediments such as loess and paleosol, as well as in lacustrine and marine sediments, have been widely used to reconstruct paleovegetation and paleoenvironments, and have provided important insights into patterns of past climate and environmental changes. For example, numerous researchers have used organic carbon isotopes of loess to reconstruct paleovegetation and paleoprecipitation. Vidic and Montañez (2004) conducted a reconstruction of paleovegetation of the central Chinese Loess Plateau during the Last Glaciation (LG) and Holocene using organic carbon isotopes in loess from Jiaodao, Shanxi Province. Hatté and Guiot (2005) carried out a paleoprecipitation reconstruction by inverse modeling using the organic carbon isotopic signal of the Nußloch loess sequence (Rhine Valley, Germany). Rao et al. (2013) reconstructed high-resolution summer precipitation variations on the western Chinese Loess Plateau during

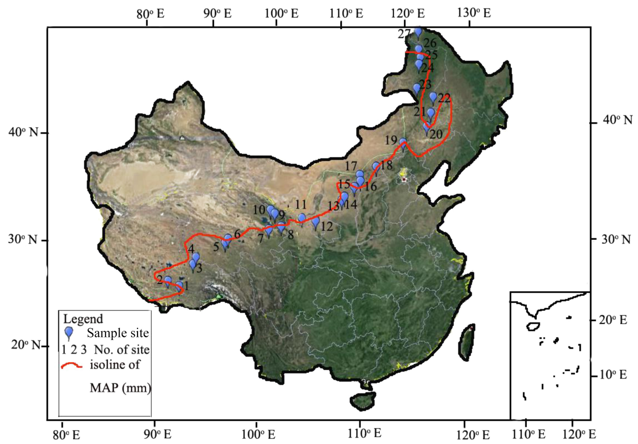

Figure 1. Sketch of the sampled region. Sample sites are indicated with numbers. 1, LangKaZi; 2, RiKaZe; 3, NaQu; 4, NieRong; 5, ZhiDuo; 6, QuMaLai; 7, TongDe; 8, TongRen; 9, HuangYuan; 10, HaiYan; 11, YuZhong; 12, XiJi; 13, JingBian; 14, HengShan; 15, ShenMu; 16, HeQu; 17, ZhunGeErQi; 18, FengZhen; 19, DuoLun; 20, LinXi; 21, ZhaLuTeQi; 22, WuLanHaoTe; 23, AErShan; 24, YaKeShi; 25, KuDuEr; 26, GenHe; 27, BeiJiCun. Detailed information of sites is shown in Table 1.

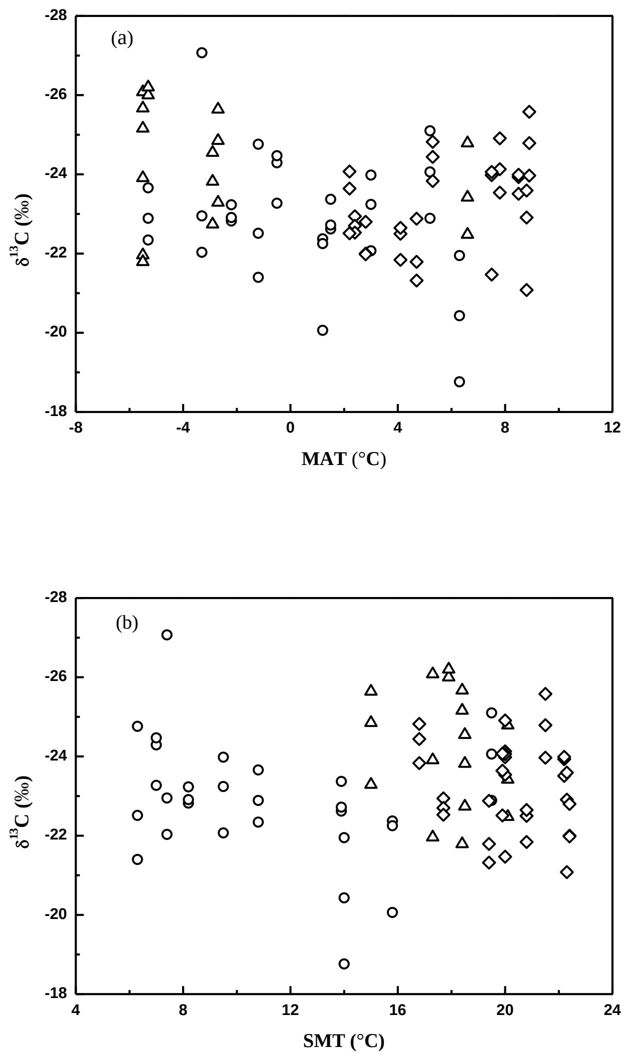

Figure 2. Shows the variance in surface soil δ^{13}C with MAT **(a)** and SMT **(b)** along the 400 mm isoline in China. Circle represents alpine and subalpine; diamond indicates temperate steppe and grassland; triangle is coniferous forest.

the LG using a well-dated organic carbon isotopic dataset. Yang et al. (2015) reconstructed a minimum 300 km northwestward migration of the monsoon rain belt from the Last Glacial Maximum to the Mid-Holocene using organic carbon isotope data from 21 loess sections across the Loess Plateau. However, to our knowledge, there are no paleotemperature reconstructions using organic carbon isotope records of loess and paleosol because it has been argued that temperature exerts only a slight or even no influence on soil organic carbon isotope values ($\delta^{13}C_{SOM}$). While this may be likely, it needs to be investigated because few studies have addressed the influence of temperature on organic carbon isotopes of modern surface soil. Lee et al. (2005) and Feng et al. (2008) both reported no relationship between temperature and surface soil δ^{13}C in central–East Asia. However, Lu et al. (2004) discovered a nonlinear relationship between mean annual temperature (MAT) and $\delta^{13}C_{SOM}$ for the Qinghai–Tibetan Plateau. Sage et al. (1999) compiled the data from Bird and Pousai (1997) and also found a nonlinear trend for the variation in $\delta^{13}C_{SOM}$ along a temperature gradient in Australian grasslands and savannas.

Plant residues are the most important source of soil organic matter. Values for $\delta^{13}C_{SOM}$ are generally close to plant δ^{13}C values, despite isotopic fractionation during decomposition of organic matter (Nadelhoffer and Fry, 1988; Balesdent et al., 1993; Ågren et al., 1996; Fernandez et al., 2003; Wynn, 2007). Thus, the factors influencing plant δ^{13}C might also influence $\delta^{13}C_{SOM}$. Plant δ^{13}C values, especially those of C_3 plants, are tightly associated with precipitation, suggesting that precipitation may also affect soil δ^{13}C (Diefendorf et al., 2010; Kohn, 2010). In addition to the effect of precipitation, numerous other factors such as temperature, air pressure, atmospheric CO_2 concentration, altitude, latitude, and longitude may also influence δ^{13}C in plants (Körner et al., 1991; Hultine and Marshall, 2000; Zhu et al., 2010; Xu et al., 2015). Although variation patterns of plant δ^{13}C with respect to temperature are so far unresolved (e.g., Schleser et al., 1999; McCarroll and Loader, 2004; Treydte et al., 2007; Wang et al., 2013), it is widely accepted that temperature has a slight effect on plant δ^{13}C. Therefore, if the ^{13}C enrichment during soil organic matter decomposition is a constant value, we expect only a slight or no influence of temperature on soil δ^{13}C. However, ^{13}C-enrichment is affected by environmental and biotic factors (Wang et al., 2015). Thus, it is difficult to determine whether or how temperature affects soil δ^{13}C, and there should be specific investigations focusing

Table 1. Information of the sampling sites.

No.	Site name	MAT/°C	SMT/°C	MAP/mm	Alt./m	Lat./N°	Long/E°	Mean $\delta^{13}C$ (‰)	Vegetation type	Dominant species	Soil types
1	LangKaZi	3.0	13.9	376	4492	29.06	90.39	−23.0	Alpine grassland	*Stipa Festuca* and *Carex*	Mattic Gelic Cambosols (Cambisols)
2	RiKaZe	6.3	14.0	420	3865	29.33	88.98	−20.4	Alpine grassland	*Stipa Festuca* and *Carex*	Mattic Gelic Cambosols (Cambisols)
3	NaQu	−3.3	7.4	406	4519	31.41	91.96	−23	Alpine meadow	*Kobresia*	Mattic Gelic Cambosols (Cambisols)
4	NieRong	−2.2	8.2	400	4731	32.09	92.27	−22.9	Alpine meadow	*Kobresia*	Mattic Gelic Cambosols (Cambisols)
5	ZhiDuo	−1.2	6.3	394	4360	33.77	95.66	−24.0	Alpine meadow	*Kobresia*	Mattic Gelic Cambosols (Cambisols)
6	QuMaLai	−5.3	10.8	391.7	4727	34.16	95.9	−24.0	Alpine meadow	*Kobresia*	Mattic Gelic Cambosols (Cambisols)
7	TongDe	−0.5	7.0	371	3258	35.27	100.64	−21.6	Subalpine grassland	*Stipa* and *Hippolytia*	Haplic Cryic Aridosolsl (Leptosols)
8	TongRen	5.2	19.5	425.7	2467	35.55	102.03	−24.0	Subalpine grassland	*Stipa* and *Hippolytia*	Haplic Cryic Aridosolsl (Leptosols)
9	HuangYuan	1.5	9.5	408.9	2725	37.02	100.8	−22.9	Subalpine grassland	*Stipa* and *Hippolytia*	Haplic Cryic Aridosolsl (Leptosols)
10	HaiYan	1.2	15.8	400	3233	36.69	101.3	−23.1	Subalpine grassland	*Stipa* and *Hippolytia*	Haplic Cryic Aridosolsl (Leptosols)
11	YuZhong	6.6	20.1	403	2361	35.78	104.05	−23.6	Temperate coniferous and broad-leaved mixed forests	*Pinustabulaeformis*	Haplic Ustic Argosols (Luvisols)
12	XiJi	5.3	16.8	400	1982	36.02	105.88	−24.4	Temperate meadow steppe	*Stipa* and *Hippolytia*	Calcic Orthic Aridosols (Calcisols)
13	JingBian	7.8	20.0	395.4	1394	37.74	108.91	−24.2	Semi-desert grasslands	*Stipa Hippolytia* and *Ajania*	Calcic Orthic Aridosols (Calcisols)
14	HengShan	8.5	22.2	397	1131	38.04	109.24	−23.8	Semi-desert grasslands	*Stipa Hippolytia* and *Ajania*	Calcic Orthic Aridosols (Calcisols)
15	ShenMu	8.9	21.5	393	1131	38.84	110.44	−24.8	Semi-desert grasslands	*Stipa Hippolytia* and *Ajania*	Calcic Orthic Aridosols (Calcisols)
16	HeQu	8.8	22.3	426	912	39.33	111.19	−22.5	Temperate meadow steppe	*Bothriochloa* and *Pennisetum*	Mottlic Calcic Orthic Aridosols (Calcisols)
17	ZhunGeErQi	7.5	20.0	400	1236	39.87	111.18	−23.2	Temperate meadow steppe	*Stipa* and *Aneuralepidium*	Mottlic Calcic Orthic Aridosols (Calcisols)
18	FengZhen	4.7	19.4	413	1236	40.45	113.19	−22.0	Temperate typical steppe	*Stipa* and *Aneuralepidium*	Typic Calcic Ustic Isohumosols (Chernozems)
19	DuoLun	2.4	17.7	407	1245	42.18	116.47	−22.7	Temperate typical steppe	*Stipa* and *Aneuralepidium*	Typic Calcic Ustic Isohumosols (Chernozems)
20	LinXi	2.2	19.9	370	928	43.44	120.08	−23.4	Temperate typical steppe	*Stipa* and *Aneuralepidium*	Typic Calcic Ustic Isohumosols (Chernozems)
21	ZhaLuTeQi	2.8	22.4	387	332	44.61	120.97	−22.3	Temperate meadow steppe	*Stipa Aneuralepidium and Filifolium*	Pachic Ustic Isohumosols (Chernozems)
22	WuLanHaoTe	4.1	20.8	416	366	46.05	121.79	−22.3	Temperate meadow steppe	*Stipa Aneuralepidium and Filifolium*	Pachic Ustic Isohumosols (Chernozems)
23	AErShan	−2.7	15.0	391	1240	47.1	119.89	−24.6	Frigid temperate coniferous forest	*Larixgmelinii* and *Betulaplatyphylla Suk*	Umbric Gelic Cambosols (Umbrisols)
24	YaKeShi	−2.9	18.5	379	634	49.33	120.97	−23.7	Frigid temperate coniferous forest	*Larixgmelinii*	Umbric Gelic Cambosols (Umbrisols)
25	KuDuEr	−5.5	17.3	402	829	49.94	121.43	−24.0	Frigid temperate coniferous forest	*Larixgmelinii* and *Betulaplatyphylla Suk*	Umbric Gelic Cambosols (Umbrisols)
26	GenHe	−5.3	17.9	424	718	50.46	121.31	−26.1	Frigid temperate coniferous forest	*Betulaplatyphylla Suk*	Umbric Gelic Cambosols (Umbrisols)
27	BeiJicun	−5.5	18.4	450.8	603	53.29	122.15	−24.2	Frigid temperate coniferous forest	*Larixgmelinii and Pinussylvestnisvar*	Umbric Gelic Cambosols (Umbrisols)

Note: MAT, SMT, MAP, Alt, Lat. and Long are the abbreviations of mean annual temperature, summer mean temperature, mean annual precipitation, altitude, latitude, longitude, respectively. The longitude, latitude and altitude of each site were from the portable GPS; MAT and MAP represent the average values of more than 30 years; SMT presents the average value of June, July and August for more than 30 years. All climatic data were from the local meteorological stations and the China Meteorological Data Sharing Service System (http://data.cma.cn/site/index.html); the soil types are based on the Chinese Soil Taxonomy and the World Reference Base – WRB (in the parentheses).

on this issue. Although the relationship between temperature and $\delta^{13}C_{SOM}$ has been investigated in the studies mentioned above, these studies were unable to effectively separate the influence of temperature from the effect of precipitation. In addition, there are no meteorological stations near most of the sampling sites in the aforementioned studies, suggesting that meteorological data had to be interpolated, which can lead to unrealistic precipitation data in regions with strong topographical variability. This interpolation could have introduced errors in the relationships between temperature and $\delta^{13}C_{SOM}$ that were established in these studies.

The present study includes a detailed investigation of the variation in $\delta^{13}C_{SOM}$ with respect to temperature across a temperature gradient along the 400 mm isohyet (isopleth of mean annual precipitation; MAP) in China. We sampled surface soil along a specific isohyet to minimize the effect of precipitation changes on $\delta^{13}C_{SOM}$.

In addition, we collected samples only at sites with meteorological stations. Thus, the climatic data we obtained from these stations are likely more reliable than interpolated values.

2 Materials and methods

2.1 Study site

In this study, we set up a transect along the 400 mm isohyet from Langkazi (site 1, 29°3.309′ N, 90°23.469′ E) on the Qinghai–Tibetan Plateau in southwest China to Beijicun (site 27, 53°17.458′ N, 122°8.752′ E) in Heilongjiang Province, northeast China (Fig. 1, Table 1). The straight-line distance between the two sites is about 6000 km. Twenty-seven (27) sampling sites were set along the transect. Among these sampling sites, 10 sites are located on the Qinghai–Tibetan Plateau and the remaining sites are in north China. Beijicun and Kuduer have the lowest MAT of −5.5 °C, while Shenmu has the highest MAT of +8.9 °C. The average MAP of these sites is 402 mm. In north China, rainfall from June to September accounts for approximately 80 % of the total annual precipitation, and the dominant control over the amount of precipitation is the strength of the East Asian monsoon system. On the Qinghai–Tibetan Plateau, however, precipitation is associated with both the Southwest monsoon and the Qinghai–Tibetan Plateau monsoon; approximately 80–90 % of rainfall occurs in the summer season (from May to October).

2.2 Soil sampling

Soil samples were collected in the summer of 2013 between 12 July and 30 August. To avoid disturbance by human activities, sample sites were chosen 5–7 km from the towns where the meteorological stations are located. We set three squares (0.5 × 0.5 m) within a 200 m^2 area to collect surface mineral soil (0–5 cm) using a ring knife. The O-horizon, including litters, moders, and mors, was removed before collecting mineral soils. About 10 g of air-dried soil was sieved using a 2 mm mesh. Plant fragments and the soil fraction coarser than 2 mm were removed. The remainder of the sieved sample was immersed in HCl (1 mol L^{-1}) for 24 h. To ensure that all carbonate was removed, the samples were stirred four times during the immersion. Then, the samples were washed to neutrality using distilled water, oven-dried at 50 °C, and ground. Carbon isotope ratios were determined using a Delta^{Plus} XP mass spectrometer (Thermo Scientific, Bremen, Germany) coupled with an elemental analyzer (FlashEA 1112; CE In-

Table 2. Shows the results from multiple regressions.

Model	R^2	Adjusted R^2	F	p value
1	0.091	0.030	1.484	0.205
2	0.374	0.273	3.690	<0.001
3	0.297	0.195	2.911	0.004
4	0.362	0.247	3.164	0.001

Note: Model-1 is the multiple regression of soil δ^{13}C against MAT, MAP, altitude, latitude and longitude; for Model-2, Model-3 and Model-4, in addition to taking these five environmental factors as independent variables, the soil types based on Chinese nomenclature and the WRB, and the vegetation types as dummy variables were separately introduced in the multiple regressions.

struments, Wigan, UK) in continuous flow mode. The elemental analyzer combustion temperature was 1020 °C.

Carbon isotopic ratios are reported in delta notation relative to the V-PDB standard using the following equation:

$$\delta^{13}\mathrm{C} = (R_{\mathrm{sample}}/R_{\mathrm{standard}} - 1) \times 1000, \tag{1}$$

where δ^{13}C is the carbon isotope ratio of the sample (‰) and R_{sample} and R_{standard} are the ^{13}C / ^{12}C ratios of the sample and the standard, respectively. We obtained a standard deviation of less than 0.15 ‰ among replicate measurements of the same soil sample.

3 Results

Except for one $\delta^{13}C_{SOM}$ value (−18.8 ‰), all other data vary between −20.4 and −27.1 ‰, with a mean value of −23.3 ‰ ($n = 80$, SD = 1.45). Multiple regressions with MAT, MAP, altitude, latitude, and longitude as independent variables and $\delta^{13}C_{SOM}$ as the dependent variable show that only 9 % of the variability in soil δ^{13}C can be explained by a linear combination of all five environmental factors ($p = 0.205$; Table 2). Considering the possibility of correlations among the five explanatory variables, stepwise regression was used to eliminate the potential influence of collinearity among them. Variables with P values < 0.05 were incorporated into the model and variables with P values > 0.1 were excluded. Statistical analysis shows that only latitude is included in the stepwise regression model ($R^2 = 0.077$, $p = 0.012$). In order to better constrain the relationship between soil δ^{13}C and each environmental factor, bivariate correlation analyses of soil δ^{13}C against some of the environmental factors were conducted. The bivariate correlation analyses show that $\delta^{13}C_{SOM}$ is not related to MAT ($p = 0.114$) or SMT ($p = 0.697$) along the isohyet (Fig. 2a, b). In addition, in order to further determine the response of $\delta^{13}C_{SOM}$ to temperature, we considered three subsets of our soil samples defined according to the climate, topography, or vegetation type of the Qinghai–Tibetan Plateau (mainly alpine meadow, 10 sites), steppe or grassland (11 sites), and coniferous forest (6 sites; Table 1). Bivariate correlation analyses within these subsets also show no relationship between $\delta^{13}C_{SOM}$ and MAT for all categories. The correlation analysis of $\delta^{13}C_{SOM}$ with respect to altitude is shown in Fig. 3, which displays no relationship ($p = 0.132$). Although longitude is not found to influence $\delta^{13}C_{SOM}$ in the above stepwise regression, bivariate correlation analyses show that both latitude and longitude are negatively correlated with $\delta^{13}C_{SOM}$ ($p = 0.012$ and 0.034, respectively; Fig. 4a, b).

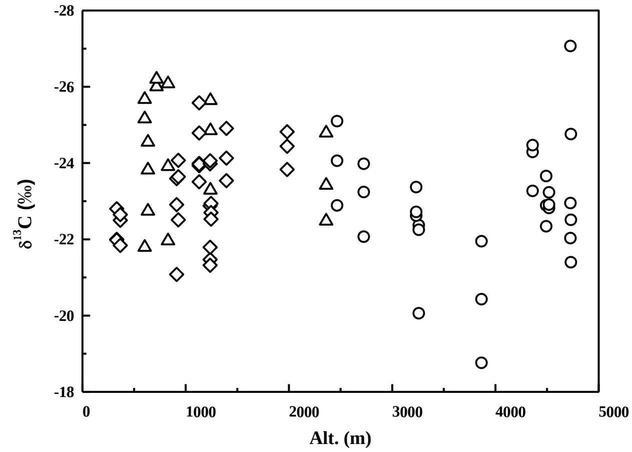

Figure 3. Shows the variance in surface soil δ^{13}C with altitude.

In addition to the effects of quantifiable environmental factors, qualitative factors such as soil type and vegetation type may influence $\delta^{13}C_{SOM}$. Various concepts have been introduced in soil taxonomy, leaving varied soil nomenclatures in use. In this study, we adopted the Chinese Soil Taxonomy and the World Reference Base (WRB) to describe the observed soils. The soil samples can be divided into eight or six types based on the Chinese Soil Taxonomy or the WRB, respectively (Table 1). One-way ANOVA analyses suggest that both soil and vegetation type play a significant role for $\delta^{13}C_{SOM}$ ($p = 0.002$ for soil type based on the Chinese Soil Taxonomy, $p = 0.003$ for soil type based on the WRB, and $p = 0.001$ for vegetation type; Fig. 5).

To further constrain the effects of soil and vegetation type on $\delta^{13}C_{SOM}$, multiple regressions with soil and vegetation type as dummy variables were conducted. Considering the tight relationship between soil type and vegetation type, especially in the Chinese Soil Taxonomy, soil variables and vegetation variables were separately introduced into the statistical analyses. Multiple regression, in which the five aforementioned explanatory environmental factors were taken as quantitative variables and the eight soil types of the Chinese nomenclature as values of a dummy variable, shows that environmental factors and soil types account for 37.4 % of the soil δ^{13}C variance ($p < 0.001$; Table 2). Using the six soil types based on the WRB rather than the Chinese nomenclature, 29.7 % ($p = 0.003$) of the variability is explained (Table 2). Similarly, multiple regression with vegetation types as dummy variables shows that the five environmental fac-

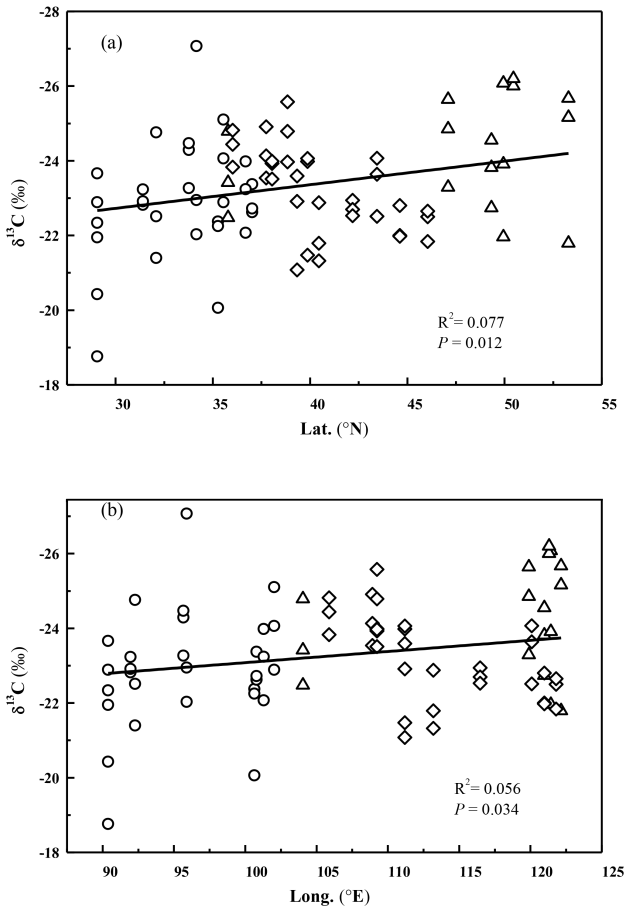

Figure 4. Shows the relationships between the soil δ^{13}C and latitude **(a)** and longitude **(b)**.

tors and vegetation types together can explain 36.2 % of the variability in soil δ^{13}C ($p = 0.001$; Table 2). Compared to the multiple regressions in which only quantitative environmental variables were introduced, the multiple regressions in which soil and vegetation were also introduced explain more of the variance, suggesting that soil type and vegetation type play a significant role in $\delta^{13}C_{SOM}$ variability.

4 Discussion

Soil δ^{13}C depends on the δ^{13}C of plants and on carbon isotopic fractionation during organic matter decomposition. δ^{13}C values of C_3 plants vary between −22 and −34 ‰ with a mean of −27 ‰, and C_4 plants range from −9 to −19 ‰ with a mean of −13 ‰ (Dienes, 1980). Carbon isotope fractionation occurs during the process of plant litter decomposition to soil organic matter in most environments, especially in non-arid environments, causing ^{13}C-enrichment in soil organic matter compared to the plant sources (Nadelhoffer, 1988; Balesdent et al., 1993; Ågren et al., 1996; Fernandezet al., 2003; Wynn et al., 2005; Wynn, 2007). A detailed

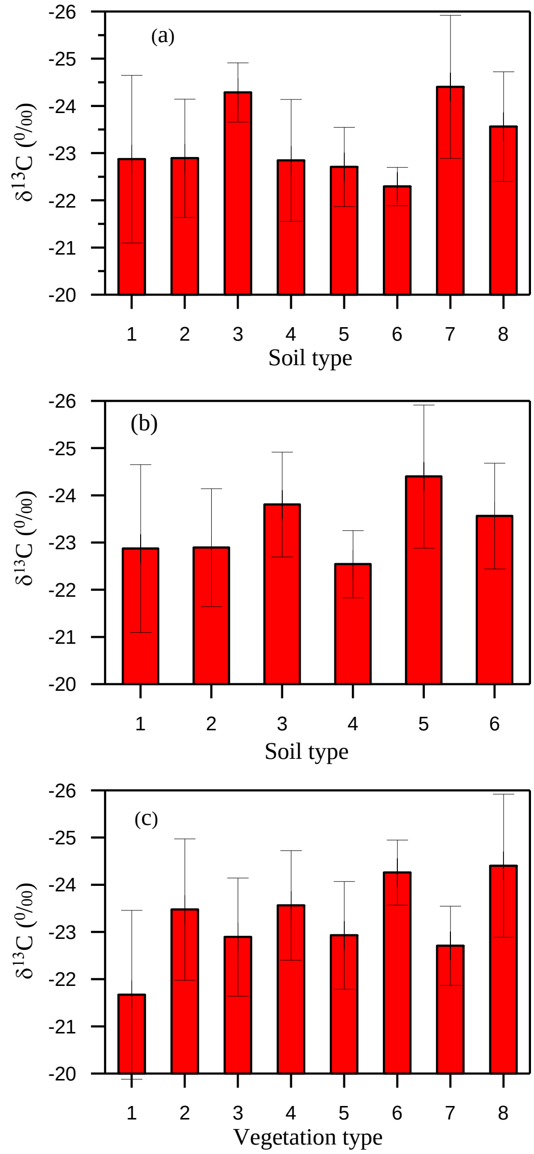

Figure 5. Shows the effects of soil types and vegetation types on the soil δ^{13}C. **(a)** Soil types based on Chinese soil taxonomy. 1. Mattic Gelic Cambosols; 2. Haplic Cryic Aridosols; 3. Calcic Orthic Aridosols; 4. Mottlic Calcic Orthic Aridosols; 5. Typic Calcic Ustic Isohumosols; 6. Pachic Ustic Isohumosols; 7. Umbric Gelic Cambosols; 8. Haplic Ustic Argosols. **(b)** Soil types based on the WRB. 1. Cambisols; 2. Leptosols; 3. Calcisols; 4. Chernozems; 5. Umbrisols; 6. Luvisols. **(c)** Vegetation type. 1. Alpine grassland; 2. alpine meadow; 3. subalpine grassland; 4. temperate coniferous and broad-leaved mixed forests; 5. temperate meadow steppe; 6. semi-desert grasslands; 7. temperate typical steppe; 8. frigid temperate coniferous forest. The bar in Fig. 5 indicates ±1 SD.

investigation of isotope fractionation during organic matter decomposition, which was conducted on Mount Gongga, an area of the Qinghai–Tibetan Plateau dominated by C_3 vegetation with herbs, shrubs, and trees, showed that the mean ^{13}C-enrichment in surface soil (0–5 cm depth) relative to the vegetation was 2.87 ‰ (Chen et al., 2010). Another investigation of 13 soil profiles from the Tibetan Plateau and north China showed that the δ^{13}C difference between surface soil and the original biomass varied from 0.6 to 3.5 ‰ with a mean of 1.8 ‰ (Wang et al., 2008). Thus, the $\delta^{13}C_{SOM}$ dataset from this study ($\delta^{13}C_{SOM}$ ranges from −20.4 to −27.1 ‰) indicates that the modern terrestrial ecosystem along the isohyet is dominated by C_3 plants. This result is consistent with the observations of vegetation along the isohyet completed in our previous study (Wang et al., 2013). The relatively heavy soil δ^{13}C values (mean: −20.4 ‰) at Rikaze (Site 2; Fig. 3 and Table 1) are surprising because only four species of C_3 plants grow there, and C_4 species are absent. This observation suggests that very large carbon isotope fractionation during SOM degradation has occurred in the local ecosystem. Previous studies have observed a similar phenomenon, although the mechanism responsible for the unusually large isotopic fractionation remains unclear. For example, Wynn (2007) reported that isotopic fractionation enriched soil organic carbon by up to ∼ 6 ‰ with respect to the original biomass.

The MAT, MAP, altitude, latitude, and longitude combined are responsible for only 9 % of the variability in soil δ^{13}C in the multiple regression model, suggesting that the contribution of these five environmental factors to soil δ^{13}C variance is very small. Our previous study conducted along the same isohyet indicated a strong positive relationship between the δ^{13}C of plants and MAT, with a coefficient of 0.104 ‰ °C^{-1} (Wang et al., 2013). The difference between the maximum and minimum temperature along the isohyet is 15 °C, so the greatest possible effect of temperature on plant δ^{13}C along the temperature gradient is 1.56 ‰, which is not very substantial. Because the main source of soil organic matter along the isohyet is C_3 plants, the induced variance in soil δ^{13}C by plant δ^{13}C can also not be very high. On the other hand, although the ^{13}C-enrichment during SOM degradation follows a Rayleigh distillation process (Wynn, 2007), our recent study shows that temperature does not influence carbon isotopic fractionation during decomposition of organic matter (Wang et al., 2015), which also explains the lack of a relationship between soil δ^{13}C and temperature. Feng et al. (2008) and Lee et al. (2005) reported no relationship between soil δ^{13}C and MAT and SMT, respectively, which is consistent with our results. Their field campaigns were conducted in central Asia, which is also dominated by C_3 plants, similar to the area along the 400 mm isohyet. This is the reason why the same pattern exists in both central Asia and in the area along the 400 mm isohyet.

Observations by Bird and Pousai (1997) and Sage et al. (1999) appear to be inconsistent with our findings; the authors found a nonlinear relationship between soil δ^{13}C and MAT in Australian grasslands. However, if they considered only soil with pure C_3 plants (MAT is below 16 °C), soil δ^{13}C and temperature were not related in Australian grasslands, which agrees with our results. Below 15 °C, the C_4 contribution to productivity in Australian grasslands is negligible, whereas above 23 °C, C_3 contribution is negligible. Between 14 and 23 °C, soil δ^{13}C is positively correlated with MAT, indicating an increase in C_4 representation with increasing MAT (Sage et al., 1999). Lu et al. (2004) also reported a nonlinear relationship between soil δ^{13}C and MAT. Similarly, if their soil data with C_4 plants are excluded from the nonlinear correlation, soil δ^{13}C is also not related to MAT (see Fig. 5b in Lu et al., 2004). Thus, the present study and the previous observations are consistent in showing that in a terrestrial ecosystem in which the vegetation is dominated by C_3 plants, temperature does not influence soil δ^{13}C variance.

Because all soil samples were taken along the 400 mm isohyet, this study shows that the contribution of precipitation to the variability in soil δ^{13}C is negligible. Although stepwise regression and correlation analysis both show a significant influence of latitude on soil δ^{13}C ($p = 0.012$; Fig. 4a), which was also described by Bird and Pausai (1997) and Tieszen et al. (1979), the five environmental variables, including latitude, are responsible for only 9 % of the variability in soil δ^{13}C in a multiple regression model (Table 2), suggesting that the contribution of latitude to soil δ^{13}C is also limited. Latitude is a comprehensive environmental factor, and change in latitude can bring about changes in other environmental factors, such as temperature, irradiation, cloud amount, and moisture. Among those, temperature and irradiation should be most strongly related to latitude. The observed relationship between latitude and soil δ^{13}C suggests that environmental factors other than temperature might also contribute to the variance in soil δ^{13}C.

Control of soil δ^{13}C by vegetation type mainly reflects the effect of plant life forms on plant δ^{13}C, which in turn influences isotope fractionation during organic matter decomposition. Communities in which life forms of dominant plants are similar are generally treated as the same vegetation type. Plant δ^{13}C is closely related to plant form (Diefendorf et al., 2010; Ehleringer and Cooper, 1988), which causes δ^{13}C differences among varying vegetation types, resulting in the observed effect of vegetation type on soil δ^{13}C.

Substrate quality partly quantifies how easily organic carbon is used by soil microbes (Poage and Feng, 2004). It can be related to plant type and is often defined using the C / N ratio, lignin content, cellulose content, and/or lignin content / N ratio (Melilloet al., 1989; Gartern et al., 2000). Our study of Mount Gongga, China, showed that litter quality plays a significant role in isotope fractionation during organic matter decomposition, and that the carbon isotope fractionation factor α increases with litter quality (Wang et al., 2015). Thus, the isotope fractionation factor should differ among sites because litter quality is dependent on vegetation, which causes changes in soil $\delta^{13}C$ with vegetation type.

The effect of soil type on soil $\delta^{13}C$ may be associated with the effect of soil type on isotope fractionation during organic matter decomposition, which involves at least two mechanisms (see Wang et al., 2008, for a detailed discussion). First, properties and compositions of microbial decomposer communities are dependent on soil type (Gelsomino et al., 1999). Different microbes can have different metabolic pathways, even when they decompose the same organic compound (Macko and Estep, 1984), and the extent of isotope fractionation during decomposition may be closely related to the metabolic pathways of microbes (Macko and Estep, 1984). Second, physical and chemical properties such as pH, particle size fraction, and water-holding capacity are considerably different among soil types, which causes organic compounds to decay at different rates in different soil environments. The magnitude of isotope fractionation during decomposition is linked to the degree of organic matter decomposition (Feng, 2002). Thus, soil type plays a significant role in soil carbon isotopic fractionation.

5 Conclusions

The present study analyzed organic carbon isotopes in surface soil along a 400 mm isohyet of mean annual precipitation in China. Our results indicate that both soil type and vegetation type significantly influence soil organic carbon isotopes. However, temperature is found to have no observable impact on $\delta^{13}C_{SOM}$, suggesting that $\delta^{13}C$ signals in sediments cannot be used for temperature reconstructions and that the effect of temperature on $\delta^{13}C_{SOM}$ should be neglected in reconstructions of paleoclimate and paleovegetation that use carbon isotopes of soil organic matter.

6 Data availability

There is no underlying material and related items in this paper. All data will be provided in the Supplement.

The Supplement related to this article is available online at doi:10.5194/bg-13-5057-2016-supplement.

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