Biogeosciences Discuss., 5, 1795–1823, 2008
www.biogeosciences-discuss.net/5/1795/2008/

Biogeosciences Discussions

Biogeosciences Discussions is the access reviewed discussion forum of *Biogeosciences*

Paleovegetation reconstruction using δ^{13}C of Soil Organic Matter

G. Wang[1], X. Feng[2], J. Han[3], L. Zhou[4], W. Tan[1], and F. Su[1]

[1]Key Laboratory of Plant-Soil Interactions, Ministry of Education, College of Resources and Environment, China Agricultural University, Beijing 100094, China
[2]Department of Earth Sciences, Dartmouth College, Hanover, NH, USA
[3]Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing 100029, China
[4]College of Environmental Sciences, MOE Laboratory for Earth Surface Processes, Peking University, Beijing 100871, China

Received: 28 February 2008 – Accepted: 12 March 2008 – Published: 18 April 2008

Correspondence to: G. Wang (gawang@cau.edu.cn)

Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

The relative contributions of C_3 and C_4 plants to vegetation at a given locality may be estimated by means of $\delta^{13}C$ of soil organic matter. This approach holds great potential for paleoecological reconstruction using paleosols. However, two uncertainties exist, which limits the accuracy of this application. One is ^{13}C enrichment as plant carbon becomes incorporated into soil organic matter. The other is due to environmental influences on $\delta^{13}C$ of plants. Two types of data were collected and analyzed with an objective of narrowing the error of paleovegetation reconstruction. First, we investigated $\delta^{13}C$ variations of 557 C_3 and 136 C_4 plants along a precipitation gradient in North China. A strong negative relationship is found between the $\delta^{13}C$ value of C_3 plants averaged for each site and the annual precipitation with a coefficient of −0.40‰/100 mm, while no significant coefficients were found for C_4 plants. Second, we measured $\delta^{13}C$ of soil organic matters for 14 soil profiles at three sites. The isotopic difference between vegetation and soil organic matter are evaluated to be 1.8‰ for the surface soil and 2.8‰ for the soil at the bottom of soil profiles. Using the new data we conducted a sample reconstruction of paleovegetation at the central Chinese Loess Plateau during the Holocene and the Last Glaciation, and conclude that, without corrections for ^{13}C enrichment by decomposition, the C_4 abundance would be overestimated. The importance and uncertainties of other corrections are also discussed.

1 Introduction

Carbon isotopic ratios of soil organic matter ($\delta^{13}C_{SOM}$) are close to that of the vegetation, and thus $\delta^{13}C_{SOM}$ can be used to estimate the relative abundance of C_3 and C_4 biomass in the vegetation at a given locality in the past. Many researchers have used $\delta^{13}C_{SOM}$ of paleosols and/or loess to reconstruct paleovegetation and paleoclimate (e.g. Stanley et al., 1991; Boutton, 1996; Boutton et al., 1998; Guillaume et al., 2001; Wang and Zheng, 1989; Gu, 1991; Frakes and Sun, 1994; Han et al., 1996; Wang and

Follmer, 1998; Lin et al., 1991; Ding and Yang, 2000; Vidic and Montañez, 2004; Liu et al., 2005; An et al., 2005). These reconstructions use isotope mass-balance equations of the following:

$$C_3(\%) = (\delta^{13}C_{SOM} - \delta^{13}C_{C_4})/(\delta^{13}C_{C_3} - \delta^{13}C_{C_4}) \times 100; \quad (1)$$

$$C_4(\%) = 100 - C_3(\%) \quad (2)$$

where $\delta^{13}C_{C_3}$, $\delta^{13}C_{C_4}$ are the mean $\delta^{13}C$ values of C_3 and C_4 plants at a given locality at the time the soil under study was developed; $C_3(\%)$ and $C_4(\%)$ are percentages of C_3 and C_4 biomass in the local vegetation. The accuracy of the reconstruction depends upon the accuracy of the knowledge on the end member values of $\delta^{13}C_{C_3}$ and $\delta^{13}C_{C_4}$. The use of these equations also assumes that $\delta^{13}C_{SOM}$ represents the $\delta^{13}C$ value of bulk local vegetation.

To acquire the $\delta^{13}C$ values for the pure C_3 and C_4 plants at the time paleosols formed, isotopic values for modern plants are first obtained. However, modern plants may be isotopically different from ancient plants due to environmental changes, such as variations in the $\delta^{13}C$ and concentration of the atmospheric CO_2, and in the amount of annual precipitation (e.g. Farquhar et al., 1982; Körner et al., 1988). Attempts have been made by several authors to correct the effects of these environmental factors before Eqs. (1) and (2) were applied (Liu et al., 2005; Chen et al., 2005).

Anther source of uncertainty comes from the assumption that $\delta^{13}C_{SOM}$ equals the $\delta^{13}C$ of vegetation. It is known that isotopic enrichment occurs during decomposition such that the soil organic matter tends to have higher $\delta^{13}C$ values than the vegetation (Troughton et al., 1974; Stout et al.,1975; Schleser and Pohling, 1980; Balesdent et al., 1993; Boutton, 1996; Bird and Pousal, 1997). To our knowledge, this effect has not been taken into consideration when Eqs. (1) and (2) are used for paleovegetation reconstruction. It has been argued that using specific biomarkers, such as n-alkanes, this source of error can be ignored (Zhang et al., 2004). While this statement is likely

but yet to be demonstrated, it is certain that isotopic enrichment in the bulk soil organic matter relatively to the vegetation due to decomposition should be corrected.

This paper evaluates various corrections described above for paleovegetation reconstructions. We particularly focus on the precipitation effect on $\delta^{13}C$ in both C_3 and C_4 plants, and the ^{13}C enrichment of soil organic matter relative to vegetation due to decomposition. We have collected new data for quantifying these two effects.

New data is necessary because the existing data do not yield certain sensitivities for these two effects. Many studies have observed that $\delta^{13}C$ of C_3 plants decreases significantly with increases in precipitation (e.g. Ehleringer and Cooper, 1988; Smedly et al., 1991; Wang et al., 2003, 2005; Schulze et al., 2006). Liu et al. (2005) measured $\delta^{13}C$ of three C_3 species occurring in Northwest China, and found that the mean $\delta^{13}C$ decrease is 1.1‰ with every 100 mm increase in annual precipitation. This sensitivity is much greater than that reported by Steward et al. (1995) (−0.34‰/100 mm) and Wang et al. (2003) (−0.49‰/100 mm). Relative to C_3 plants, the $\delta^{13}C$ of C_4 plants is not sensitive to water availability (Farquhar et al., 1982; Henderson et al., 1992; Wang et al., 2005, 2006). Studies on C_4 species mostly showed slight increases in $\delta^{13}C$ with increasing water availability (Buchmann et al., 1996; Schulze et al., 1996; Wang et al., 2005, 2006). However, Liu et al. (2005) observed an opposite pattern in *Bothriochloa ischaemum* (C_4), i.e., $\delta^{13}C$ decreasing significantly with increasing precipitation (−0.61‰/100 mm). Our study includes an intensive investigation on the plants' $\delta^{13}C$ response to precipitation in an arid and semiarid region by averaging a large number of species in 33 sites in North China with annual precipitation ranging from 15 to 650 mm.

The ^{13}C enrichment in soil organic matter has been observed in both field and laboratory investigations. Some observations have come from field or laboratory incubation experiments (e.g. Natelhoffer and Fry, 1988; Wedin et al., 1995; Fernandez et al., 2003; Connin et al., 2001). Time spans used in these experiments are relatively short, typically less than 5 years, while the time for decomposition of organic matters in natural systems is much longer, on timescales from decades to millennia. Therefore, these short-term experimental results represented ^{13}C enrichments only during

the early stage of decomposition. In fact, progressive ^{13}C enrichment continues even during decomposition of resistant soil organic matter having slow turnover rates (Chen et al., 2002; Feng et al., 1999; Torn et al., 2002). Other observations have come from isotopic measurements of soil profiles (e.g. Stout et al., 1978; Dzurec et al., 1985; Gregorich et al., 1995; Balesdent et al., 1993; Boutton, 1996; Bowman et al., 2002). Most of such studies aimed to understand vegetation dynamics, and thus study sites were chosen in areas where C_3/ C_4 ratios in local vegetation have not been constant in the past (e.g. Dzurec et al., 1985; Schwartz et al., 1986; Boutton et al., 1998; Guillaume et al., 2001; Krull et al., 2005). As a result, the observed $\delta^{13}C$ differences between soil organic matter and surface standing vegetation included not only the isotopic enrichment during organic matter decomposition, but also changes of the C_3/C_4 ratio in the local vegetation. To quantify isotopic enrichment in soil organic matter, we studied isotopic variations in 14 soil profiles at three sites. All three sites are undisturbed, and have C_3 only vegetations; they are chosen especially because we intended to avoid isotopic variations in soil organic matter due to changing abundance of C_3 and C_4 plant types. We then evaluate how various corrections affect the paleovegetation reconstruction for Chinese Loess Plateau.

2 Study sites and methods

2.1 Study sites of plant sampling and plant collection

A total of 33 sites in North China were studied, among which 27 sites are in Northwest China (Fig. 1). From west to east, North China is characterized by temperate arid, semiarid and semi-humid climate. The dominant control over the precipitation amount is the strength of the East Asian summer monsoon system. The precipitation occurs mostly in the summer season (from May to September), which accounts for approximately 68% to 87% of the total annual precipitation. From southeast to northwest in the study region, the annual rainfall amount decreases from 650 to 15 mm. The vegetation

is dominated by shrubs and herbs.

Plants were sampled in August of 1998 to 2005. In order to minimize the influences attributed to human disturbance, local water supply, light regime or location within the canopy, sampling was restricted to unshaded sites far from human habitats. All the species collected were those having high abundance in the community or those widely occurring in North China At each site, 5–7 plants of each species of interest were identified, and the uppermost and the second fully opened leaves of each individual were obtained. The leaves from each species at each site were mixed together into one sample. A total of 557 C_3 samples (including 255 species) and 136 C_4 samples (including 43 species) were collected.

2.2 Study sites of soil sampling and soil collection

One of the three soil sampling sites (Site 1) is located at the northeastern part of the Qinghai-Tibetan Plateau in which the Chinese Academy of Sciences has a research station, Haibei Highland Frigid Meadow Ecosystem Experimental Station (101°12′E, 37°45′N; 3200 m a.s.l.). The other two sites (Sites 2 and 3) are in the Donglingshan Mountain situated on the boundary between Beijing and Hebei Province in North China, approximately 130 km west of Beijing city (Fig. 1). These two sites are within Beijing Forest Ecosystem Research Station, which covers 16 000 ha of land, and was established on Donglingshan Mountain by the Chinese Academy of Sciences in 1990.

Site 1 has the characteristics of a highland continental climate, and is cold and humid with the mean annual temperature of −2.0°C and the mean annual precipitation of around 600 mm. Vegetation there is highland frigid meadow with most dominant species of *Potentilla fruticosa*, *Kobresia pygaea* and *Kobresia humilis*. There are no C_4 species. Soil type is alpine meadow soil. The area of Sites 2 and 3 is characterized by temperate semimoist climate. The amount of annual precipitation is 611.9 mm and the mean annual temperature is from 2°C (2303 m a.s.l. at the peak of Donglingshan Mountain) to 8°C (400 m a.s.l., Zaitang). The vegetation is characterized by forest, shrub and subalpine meadow. Site 2 is situated on the north-facing slope of Dongling-

shan Mountain at an elevation of 1600 m. The vegetation is a deciduous broad-leaved forest with only one birch tree species, *Betula platyphylla,* having an average height of 5 m and an average canopy area of 10 m^2. In addition, two C_3 grass species, *Trullius chinensis* and *Galium verum*, with about 0.15 m and 0.10 m high, respectively, occur within birch stands. No C_4 species is present at this site. The site has a brown forest soil. Site 3 is located on a ridge with an elevation of 1700 m, approximately 300 m southeast of Site 2. The vegetation at Site 3 is a subalpine meadow with *Roegneria kamoji* and *Artemisia lavandulaefolia* grasses as dominant species. Except for few *Setaria viridis* (C_4) occurring on roadsides and neighbor of the climatic station, which is also located at the ridge (1700 m a.s.l.), no C_4 species exists elsewhere, particularly no C_4 species at the sampling plots. Soil type is subalpine meadow soil. There is no human habitat within 20 km^2 of Site 2 and Site 3, mainly because of strong winds and low temperature for many days of each year. Grazing and cutting have been strictly prohibited in the area since 1980's.

At Site 1, six 0.5 m×0.5 m plots were laid out within one of the enclosures in Haibei Highland Frigid Meadow Ecosystem Experimental Station with *Kobresia pygaea* and *Kobresia humilis* as dominant species. For each plot, all aboveground plants were collected (and will be used as litter input to the soil), and then the soil profile was dug to the weathered rock (saprolite). Soil profiles were sampled at 5 cm intervals. The depths of the six soil profiles are 60 cm, 120 cm, 110 cm, 60 cm, 70 cm and 90 cm, respectively.

At Site 2, five plots of 0.5 m×0.5 m were sampled in *Betula platyphylla* stands. In order to avoid birch roots, all plots were chosen at 2–3 m away from tree trunks. At each plot, we first sampled the litter layer, and then dug a soil profile to the weathered rock. The depths of the five soil profiles are 75 cm, 105 cm, 100 cm, 100 cm and 120 cm, respectively. All soil profiles were sampled at 5 cm intervals except for one that were sampled at 10 cm intervals.

At Site 3, three plots of 0.5 m×0.5 m were sampled in the same way as described for Site 1. The depths of the soil profiles are 60 cm, 70 cm and 95 cm, respectively, and

soils were sampled at 5 cm intervals.

2.3 Measurements

The plant and litter samples were oven-dried, and ground to 40 mesh. Soil samples were air-dried and sieved with a 2-mm sieve. Plant fragments and the soil fraction coarser than 2 mm were removed. The <2 mm soil fraction was treated with 1N HCl at room temperature overnight to remove carbonates, after which it was washed and oven-dried at 70°C. Carbon contents and carbon isotopic ratios of samples were determined on an elemental analyzer coupled with a Delta ^{Plus} XP mass spectrometer in a continuous flow mode. The ^{13}C to ^{12}C ratio of the sample is reported in the δ-notation as the relative difference in parts per thousand (per mil) from the PDB standard. The repeatability of C contents is better than 0.1% (1σ) and that of $\delta^{13}C$ is better than 0.15‰ (1σ).

3 Results

3.1 $\delta^{13}C$ variations of C_3 and C_4 plants with annual precipitation

Values of $\delta^{13}C$ in C_3 plants we measured range from −21.7‰ to −30‰ with a mean value of −26.7‰ (n=557, s.d.=1.56), and those of C_4 plants vary from −10.0‰ to −15.8‰ with a mean of −12.8‰ (n=136, s.d.=1.06).

Fig. 2a plots the site-averaged $\delta^{13}C$ values for C_3 plants as a function of precipitation, showing that $\delta^{13}C$ increases significantly with decreasing precipitation with a slope of −0.004‰/mm (0.001 s.e., t-test, P<0.0001). C_4 plant $\delta^{13}C$ values display a slight increase trend with increasing annual precipitation with a slope of 0.001‰/mm (0.001 s.e., t-test, P=0.220), although the correlation is not significant (Fig. 2b). *Setaria vidis* (an annual grass with a stem length of 10–110 cm) is one of the most common C_4 species in China (Institute of Botany, CAS, 1987). Using this species alone, the $\delta^{13}C$

also slightly increases with precipitation, with a slope of 0.001‰/mm (0.001 s.e.t-test, P=0.321), but again the correlation is not significant (Fig. 2c).

3.2 Variations in δ^{13}C in soil profiles

In soil, the carbon contents drop rapidly with depths near the surface, and then level off and slowly approach zero in deeper layers. All three sites share this characteristic carbon distribution with depth. One example is given for Site 1 in Fig. 3a.

The carbon isotopic ratio in the soil organic matter at Site 1 increases with depth (Fig. 3b). Compared to the carbon contents (Fig. 3a), changes of δ^{13}C with depth are more gradual. The isotopic enrichments with depth do not seem to approach a constant value within the depth of observation. Carbon isotope profiles in Site 2 and Site 3 (Figs. 4a and 4b) also show the typical pattern of ^{13}C enrichment with depth.

We are mostly interested in the isotopic difference between vegetation and soil organic matter. Obviously, this difference depends upon which depth of soil is under discussion. When dealing with a paleosol, the specific paleo-soil horizon it is not always clear. We, therefore, tabulated in Table 1 observed δ^{13}C differences between vegetation (Sites 1 and 3) or litter (Site 2) and the soil organic matter either near the surface (0~5 cm of mineral soil) or at the bottom of soil profiles. Between the surface soil and vegetation, the δ^{13}C difference ranges from 0.6 to 2.5‰ (Sites 1 and 3), and between the bottom soil and vegetation, the δ^{13}C difference ranges from 2.6 to 5.9‰. At the forested Site 2, the δ^{13}C difference is 1.5–3.0‰ between the surface soil and litter, and 3.7–5.0‰ between bottom soil and litter (Table 1). Typically, the δ^{13}C of litter on the forest floor is about 0.5‰ higher than the foliage input (e.g. Stout et al., 1978; Dzurec et al., 1985; Gregorich et al., 1995; Balesdent et al., 1993; Wang, 2003). If this enrichment were taken in to account, the δ^{13}C difference between soil and vegetation would be 2.0–3.5‰ for the surface soil and 4.2–5.5‰ for the bottom soil.

4 Discussion

4.1 δ^{13}C variations of plants

For both C_3 and C_4 plants, water deficit usually causes plants to close stomatal pores to reduce water loss by transpiration, and thus the stomatal conductance (g) decreases. As a result, the ratio of intercellular to ambient CO_2 concentration (c_i/c_a) decreases. For C_3 plants, a decrease in c_i/c_a directly results in an increase in its carbon isotopic ratio (Farquhar et al., 1982). Observations of C_3 plant δ^{13}C values decreasing along a precipitation gradient reported by this work and others (e.g. Steward et al., 1995; Wang et al., 2005; Liu et al., 2005; Schulze et al., 2006) are consistent with the expectation. For C_4 plants, a decrease in the c_i/c_a ratio may lead to two opposite isotopic responses; the δ^{13}C may increase or decrease depending on the degree of leakiness, ϕ (Farquhar, 1983), the proportion of carbon dioxide produced within bundle sheath cells from C_4 acids that is not fixed by Rubisco but leaked back to mesophyll cells. According to the model developed by Farquhar (1983) (modified by Henderson et al., 1992), if the ϕ value is less than 0.35, the δ^{13}C of C_4 species decreases with decreasing c_i/c_a ratio; otherwise, δ^{13}C increases. The ϕ parameter seems to remain relatively constant for a given species under a wide range of environmental conditions (e.g., temperature, moisture and light conditions). Henderson et al. (1992) found that the ϕ values measured in 10 C_4 species, using a gas exchange method, remained around 0.21 over a range of irradiance and leaf temperature. In this and other studies (Schulze et al., 1996; Buchmann et al., 1996; Ghannoum et al., 2002; Wang et al., 2005), δ^{13}C values of C_4 species were observed to decrease with increasing water stress, corresponding to ϕ values less than 0.35. Therefore for the observation on *Bothriochloa ischaemum* by Liu et al. (2005) where the plant δ^{13}C significantly decreased with increasing annual precipitation amount, it is possible that the ϕ value of this species is greater than 0.35. *Bothriochloa ischaemum* seems to be a unique case, since it is the only C_4 species reported to have a negative δ^{13}C-precipitation correlation.

4.2 Variations in δ^{13}C and carbon contents of soil organic matter

The soil profiles of carbon contents and δ^{13}C show very different characteristics. The variations of carbon contents with depths are very similar within all profiles at Sites 1to 3 (Figs. 3a), which are similar to thousands of reported carbon distributions in soil profiles (Zinke et al., 1986; 1998). Typically, carbon contents decreases rapidly near the surface, and then slowly at deeper soil. Judging from the carbon content profile alone it seems that little soil processes is occurring at depth. Compared to the carbon contents, the profile of soil δ^{13}C also shows relatively rapid change near surface (although less than that of carbon contents), but the ^{13}C enrichment continues in deeper layers. This suggests that even slow decomposition of resistant organic matter is associated with isotopic enrichment.

Several mechanisms have been proposed to account for ^{13}C-enrichment of soil organic matter relative to original plants. The first mechanism is the decrease in the δ^{13}C of atmospheric CO_2 since the industrial revolution because of combustion of ^{13}C-depleted fossil fuels, and from the pre-industrial era to ~11 000 ^{14}C yr BP, the δ^{13}C of atmospheric CO_2 was relatively constant, with a mean of −6.40‰ (Marino et al., 1992). This mechanism is responsible for an isotopic difference of about 1.3‰ between the pre-industrial organic carbon and the present vegetation (Marino et al., 1992; Leuenberger et al., 1992). The second mechanism involves different decay rates of various components in organic matter having different δ^{13}C. These effects also significantly alter the total δ^{13}C, but the expected magnitude and direction depends on the relative proportions of components, and are not completely known (Feng, 2002). The third mechanism is isotopic fractionation during decomposition. The most important processes are microbial respiration and fermentation, leading to ^{13}C-enrichments in microbial products compared to the organic substrate (Macko and Estep, 1984; Poage and Feng 2004). This mechanism is considered to be the main reason for observed ^{13}C enrichment between litter and vegetation and with increasing soil depth.

The soil organic matter at the deepest depth of our soil profiles are likely derived

from plant materials not affected by fossil fuel combustion, because the age of soil organic matter in deep soil can be thousands of years old (Shen, 2000). We, therefore, subtract 1.3 from the observed mean δ^{13}C difference between deep soil and vegetation to account for the effect of the δ^{13}C decrease in atmospheric CO_2 since the industrial revolution. The δ^{13}C values of soil organic matter at 5 cm depth, may also have been affected by the δ^{13}C decrease in atmospheric CO_2 in recent decades. However, we did not measure the age of the soil organic matter, and thus cannot determine how much correction would be needed for the surface soil. We will use the observed difference without a correction for later discussions, but note here that this difference may be slightly overestimated.

Summarizing the above discussions, we report that the δ^{13}C difference is 1.8±0.3‰ (1 s.e., n=14) between the surface soil and vegetation, and 2.8±0.3‰ (1 s.e., n=14) between the bottom soil and vegetation. If all values in isotopic differences between soil and vegetation are pooled, the difference is normally distributed with a mean of 2.3‰ and standard deviation of 1.1‰ (n=28).

Previous investigations (e.g. Stout et al., 1978; Dzurec et al., 1985; Gregorich et al., 1995; Balesdent et al. 1993) showed that the δ^{13}C difference between vegetation and the topsoil (down to 20 cm depth) varied between 0.5‰ and 2.5‰. Wang (2001) obtained similar results from 36 soil profiles in Northwest China, and the mean δ^{13}C difference between vegetation and the topsoil layer is 2.2‰. These δ^{13}C differences obtained by previous studies were not corrected for the effect of the δ^{13}C decrease in atmospheric CO_2. Therefore, the 2.8‰ carbon isotopic difference we obtained between the bottom organic matters and vegetation is probably the maximum ^{13}C enrichment due to organic matter decomposition, because most paleosols may not represent the soil at deepest horizons. If the δ^{13}C difference of 1.8‰ between the surface soil and vegetation are considered to be the most likely value for isotopic correction due to organic matter decay, Eq. (1) should be modified to the following:

$$C_3(\%) = [(\delta^{13}C_{SOM} - 1.8 - \delta^{13}C_{C_4})/(\delta^{13}C_{C_3} - \delta^{13}C_{C_4})] \times 100 \quad (3)$$

4.3 Reconstructions of paleovegetation

Equations (1) and (3) are mass balance equations for estimating C_3 plants' abundance in the vegetation using the measured $\delta^{13}C$ of paleosol organic matter without or with a correction for isotopic enrichment due to soil organic matter decomposition. Before using these equations, however, we also need to obtain the end member $\delta^{13}C$ values of C_3 and C_4 plants at the time when the plants grew. If we start with modern isotopic compositions of C_3 and C_4 plants, we have to make several corrections to obtain the $\delta^{13}C_{C_4}$ and $\delta^{13}C_{C_3}$ values for the time in the geological history, e.g., Holocene or the Last Glaciation. These corrections account for the effects on plant $\delta^{13}C$ of 1) changes in precipitation, 2) changes in the $\delta^{13}C$ of atmospheric CO_2, and 3) changes in the CO_2 concentration in the atmosphere. Here we use the published loess $\delta^{13}C$ data of Holocene and the Last Glaciation (LG) in Chinese Loess Plateau reported by Vidic and Montañez (2004) to show how different types of corrections affect the qualitative reconstruction of paleovegetation.

First, we obtain plant $\delta^{13}C$ values in Holocene. The study site in Vidic and Montañez (2004) is located at Jiaodao close to Yanan city on the central Chinese Loess Plateau, which presently has a temperate semimoist climate with an annual precipitation amount of about 600 mm. Our study shows the mean $\delta^{13}C$ values of standing C_3 and C_4 plants in this area are −27.5‰ and −12.5‰ respectively. For Holocene, we could assume that precipitation was similar to the current level, and therefore no need to correct for the climate effect on plant $\delta^{13}C$ value. The $\delta^{13}C$ value of the atmospheric CO_2 in most part of Holocene was 1.3‰ higher than it is today (Marino et al., 1992; Leuenberger et al., 1992). Thus, $\delta^{13}C$ values of C_3, C_4 plants during Holocene would be −26.2‰ and −11.2‰ respectively (Table 2).

Without the anthropogenic input of fossil fuel CO_2, the atmospheric CO_2 concentration before the industrial resolution was 80 ppm lower than it is today. It is possible that the CO_2 concentration level also affect the $\delta^{13}C$ values of C_3 plants. However, the magnitude of this effect is not entirely clear. Polley et al. (1993) showed that C_3

plants, growing over a range of CO_2 concentrations characteristic of the Last Glacial Maximum to the present atmosphere, tended to have constant c_i/c_a ratios, suggesting a constant carbon isotope discrimination (Δ). Feng and Epstein (1995), on the other hand, reported that the mean Δ value of tree rings in four trees is positively correlated to the CO_2 concentration with a mean slope of 0.02‰/ppm. We have recently conducted a detailed study on the Δ response of ∼50 tree-ring series from widely distributed locations to atmospheric CO_2concentration in the past 100–200 years. We found that the pattern of the $\delta^{13}C$ response to the CO_2 concentration is complex. Before 1950, the mean Δ value tended to increase with the CO_2 concentration increase, while since 1950 Δ have decreased. The instantaneous rate of Δ change with the CO_2 concentration varied systematically, ranging from 0.025‰/ppm in 1850 to −0.013‰/ppm in 1995 (Wang and Feng, 2008[1]). Apparently, the $\delta^{13}C$ response to the CO_2 concentration is not linear during the past 100 years, and it is likely that this response not only depends upon the absolute level of CO_2 but also upon the rate of CO_2 concentration change. Thus, it is difficult to determine whether corrections for a CO_2 concentration change is necessary, and if so how much correction should be made. Here we will make paleoecolgoical reconstruction considering the two situations, with and without a CO_2 concentration correction. Under the situation with the correction, we use the coefficient of $\delta^{13}C$-CO_2 concentration of Feng and Epstein (1995) (−0.02‰/ppm), which has already been used by several previous authors. $\delta^{13}C$ value of C_3 plants would then be −24.6‰ in the Holocene after a 1.6‰ correction for 80 ppm change in the CO_2 concentration of the atmosphere. For C_4 plants, $\delta^{13}C$ has a lower sensitivity to CO_2 concentration change (Henderson et al., 1992), so it remains −11.2‰ (Table 2).

We now obtain $\delta^{13}C_{C_4}$ and $\delta^{13}C_{C_3}$ values for the LG. Atmospheric CO_2 concentration was 80 ppm lower and $\delta^{13}C$ value of CO_2 was 0.3‰ lower than that of Holocene (Marino et al., 1992; Leuenberger et al., 1992). With no correction for the effect of CO_2

[1] Wang, G. A. and Feng, X.: Long-term response in carbon isotope discrimination of tree rings to the atmospheric CO_2 concentration, in preparation, 2008.

concentration, the $\delta^{13}C$ values of C_3 and C_4 plants in the last ice age become −26.5‰ and −11.5‰, respectively. With corrections for the effect of CO_2 concentration, $\delta^{13}C$ values of C_3 plants in the last ice age become −23.3‰. $\delta^{13}C$ value of C_4 plants is still −11.5‰.

In addition, the precipitation at the location during the Last Glaciation was about 200 mm lower than that in Holocene (Wu et al., 1995). We correct the precipitation effect on the $\delta^{13}C$ of C_3 plants using the $\delta^{13}C$–precipitation coefficient of −0.4‰/100 mm obtained in this study, which yields a 0.8‰ correction for C_3 plants. Therefore, the $\delta^{13}C_{C_3}$ of vegetation becomes −25.7‰ without the correction for the CO_2 concentration, and −22.5‰ with the CO_2 correction. The precipitation change does not significantly affect the $\delta^{13}C_{C_4}$ value, and thus no further corrections are made for the $\delta^{13}C$ of C_4 plants. Values for various corrections are listed in Table 2.

The average $\delta^{13}C_{SOM}$ value of S_0 (Holocene paleosol) is reported to be −19.5‰ and that of L_1 (loess from the last glacial period) −22.4‰ (Vidic and Montañez, 2004). Table 2 shows the effect of each type of corrections on the estimate of C_4 plant abundance in the vegetation of a given age. One important comparison is shown by the difference using Eq. (1) versus Eq. (3) (column 4 and 5). Other things being equal, a correction for isotopic enrichment due to organic matter decomposition yields a C_4 plant abundance in the vegetation about 12% lower than that obtained without such a correction.

The correction for CO_2 concentration causes additional decrease in the estimated C_4 plant abundances. In Holocene, the CO_2 concentration correction reduces the estimated C_4 abundances by about 6–8%. The effect is more substantial for the LG; the estimated C_4 abundances become 0.9% using (1) but negative using Eq. (3). The negative value indicates that we may have overcorrected either for the enrichment of soil organic matter, or the $\delta^{13}C$ end member of C_3 plants, or both. Since our correction for organic matter decomposition is rather conservative (we used the difference between vegetation and surface soil), an overcorrection for the CO_2 concentration effect is likely

the cause. As discussed earlier, our recent compilation on tree-ring $\delta^{13}C$ series does not yield a mean sensitivity of −0.02‰/ppm. The highest sensitivity, −1.3‰/ppm, was obtained for the most recent year of 1995 in the series, and before 1995 it was consistently lower. Nevertheless, our calculations indicate that the C_4 abundance was probably very low and close to zero during the LG.

Compared to previously published paleovegetation reconstructions the estimated C_4 abundances from this study are consistently lower. Vidic and Montañez (2004) reported 53% for C_4% in Holocene, and 34% in the LG. Liu et al. (2005) also reported 53% of C_4 vegetation in Holocene, but 11% in the LG.

The differences in the percentage of C_4 plants between this study and those by Vidic and Montañez (2004) and Liu et al. (2005) come from the two sources. First, both Vidic and Montañez (2004) and Liu et al. (2005) did not take into account the ^{13}C enrichment during decomposition of organic matters (1.8‰ used in this study). Second, their corrections to obtain end members of $\delta^{13}C_{C_3}$ and $\delta^{13}C_{C_4}$ are difference from ours. Vidic and Montañez (2004) did not make any correction, and therefore they obtained 53% and 34% C_4 contribution in Holocene and the LG, respectively (see Table 2). When estimating the C_4% in the Last Glaciation, Liu et al. (2005), did corrections on the precipitation effect on plant $\delta^{13}C$ for both C_3 and C_4 plants −1.1‰/100 mm, −0.61‰/100 mm for C_3, C_4 plants, respectively). We think that the sensitivity they used for C_3 plants might have been too high considering that our study is more comprehensive. In addition, their observation on the C_4 plant response to precipitation based on one C_4 species (*Bothriochloa ischaemum*) may have been an exception rather than the rule. While Liu et al. (2005) did correct for the effect of CO_2 concentration, they overlooked the fact that the $\delta^{13}C$ value of CO_2 was 1.3‰ higher in Holocene than it is of today.

It is possible that Vidic and Montañez (2004) and perhaps also Liu et al. (2005) overestimated the percentages of C_4 plants. For example, the percentages of C_4 plants in the last ice age obtained by Vidic and Montañez (2004) reaches 34%, while we consider it unlikely that so much C_4 plants occurred during the LG when the annual

temperature was expected to have dropped by 8–10°C compared to Holocene in the China Loess Plateau (Wu et al., 1995; Ganopolski et al., 1998). The annual temperature today in Yanan is around 9°C. If the annual temperature in Holocene is close to this value, the annual temperature during the last glacial time might have approached 0°C, close to the mean annual temperature at the peak of Donglingshan Mountain in North China (2303 m a.s.l. the annual temperature is around 0°C) (Fig. 1), where no C_4 species are present above 1750 m (Liu, 2003). In addition, an investigation in Northwest China (including the Loess Plateau) showed that C_4 species in natural vegetation are rarely found at sites with annual temperature less than 3°C (Wang, 2001). Thus, we argue that the C_4 plants abundance during the Last Glaciation were likely to be close to zero.

5 Conclusions

The ^{13}C enrichment by decomposition of organic matters can result in overestimation of the percentage of C_4 plants in paleovegetation using $\delta^{13}C$ values of soil organic matter. This effect has been overlooked by previous investigators. This work provides data for quantitative corrections of the decomposition effect by measuring the $\delta^{13}C$ variation in soil organic matter with depth in 14 soil profiles from three sites. These sites all have C_3 only vegetation, and include both forest and grassland vegetation type. We report that the average $\delta^{13}C$ difference between soil organic matter and vegetation is 1.8 for the surface soil and 2.8 for the bottom (of the profiles) soil.

The effect of precipitation on $\delta^{13}C$ of C_3 and C_4 plants has been assessed by a comprehensive investigation of a large number of species (255 of C_3 and 43 C_4 species) from 33 sites in North China, adding additional data to the exiting database for simiarid to simimoist ecosystems. A strong negative relationship is found between the site-averaged $\delta^{13}C$ of C_3 plants and the annual rainfall amount with a coefficient of −0.40‰/100 mm, while no significant correlations are observed for both C_4 plants as a group and *Setaria viridis* as a single C_4 species, although $\delta^{13}C$ trends toward slight

increase with increasing precipitation.

This study demonstrates that the paleovegetation reconstruction can be significantly affected by various corrections to the $\delta^{13}C_{SOM}$ value, and to the end member $\delta^{13}C$ values of C_3 and C_4 plants before using mass balance calculations. We recommend that corrections for the changes in the atmospheric $\delta^{13}C$, ^{13}C enrichment by decomposition, and changes in precipitation at the given location are necessary. Corrections for changes in the atmospheric CO_2 concentration may also be needed, but we currently have relatively poor knowledge on how much that correction should be. Using these corrections, we show that C_4 plants were likely to be absent during the Last Glaciation on the central Chinese Loess Plateau, which, considering the paleo-temperature and precipitation condition, is consistent with distributions of C_4 plants in the modern environment.

Acknowledgements. This research was supported by grants from the National Natural Science Foundation of China (Grant No. 40673017) and Program for Changjiang Scholars and Innovative Research Team in University (No. IRT0511). We'd like to thank M. Yan for analyzing stable carbon isotope ratios in the Isotope Lab at the College of Resources and Environment, China Agricultural University.

References

An, Z. S., Huang, Y. S., Guo, Z. H., Clemens, S., Li, L., Prell, W., Ning, Y. F., Cai, Y. J., Zhou, W. J., Lin, B. H., Zhang, Q. L., Cao, Y. N., Qiang, X. K., Chang, H., and Wu, Z. K.: Multiple expansions of C_4 plant biomass in East Asia since 7 Ma coupled with strengthened monsoon circulation, Geology, 33, 9, 705–708, 2005.

Balesdent, J., Girardin, C., and Mariotti, A.: Site-related $\delta^{13}C$ of tree leaves and soil organic matter in a temperate forest, Ecology, 74, 6, 1713–1721, 1993.

Bird, M. I. and Pousai, P.: Variations of $\delta^{13}C$ in the surface soil organic carbon pool, Global Biogeochem. Cy., 112, 313–322, 1997.

Boutton, T. W., Archer, S. R., Midwood, A. J., Zitzer, S. F., and Bol, R.: $\delta^{13}C$ values of soil

organic carbon and their use in documenting vegetation change in a subtropical savanna ecosystem, Geoderma, 82, 1, 5–14, 1998.
Boutton, T. W.: Stable carbon isotope ratios of soil organic matter and their use as indicators of vegetation and climate change, in: Mass Spectrometry of soils, edited by: Boutton, T. W., Yamasaki, S. I., Dekker, M., Inc., New York, 47–82, 1996.
Bowman, D. M. J. S. and Cook, G. D.: Can stable carbon isotopes (δ^{13}C) in soil carbon be used to describe the dynamics of *Eucalyptus* savanna-rainforest boundaries in the Australian monsoon tropics?, Austral. Ecology, 27, 1, 94–102, 2002.
Buchmann, N., Brooks, J. R., Rapp, K. D., and Ehleringer, J. R.: Carbon isotope composition of C_4 grasses is influenced by light and water supply, Plant Cell Environ., 19, 4, 392–402, 1996.
Chen, Q. Q., Shen, C. D., Peng, S. L., Yi, W. X., Sun, Y. M., Li, Z. A., and Jiang, M. T.: Characteristics and controlling factors of soil organic matter turnover processes in the Subtropical mountainous area, South China. Acta Ecologia Sinica, 22, 9, 1446–1454, 2002 (in Chinese with English abstract).
Connin, S. L., Feng, X., and Virginia, R. A.: Isotopic discrimination during long-term decomposition in an arid land ecosystem, Soil Biol. Biochem., 33, 41–51, 2001.
Ding, Z. L. and Yang, S. L.: C_3/ C_4 vegetation evolution over the last 7.0 Myr in Chinese Loess Plateau: evidence from pedogenic carbonate δ^{13}C, Palaeogeogr. Palaeocl., 160, 4, 291–299, 2000.
Dzurec, R. S., Boutton, T. W., Caldwell, M. M., and Smith, B. N.: Carbon isotope ratios of soil organic matter and their use in assessing community composition changes in Curlew Valley, Utah, Oecologia, 66, 1, 17–24, 1985.
Ehleringer, J. R. and Cooper, T. A.: Correlations between carbon isotope ratio and microhabitat in desert plants, Oecologia, 76, 5, 562–566, 1988.
Farquhar, G. D.: On the nature of carbon isotope discrimination in C_4 species, Aust. J. Plant. Physiol., 10, 2, 205–226, 1983.
Farquhar, G. D, O'Leary, M. H., and Berry, J. A.: On the relationship between carbon isotope discrimination and the intercellular carbon dioxide concentration in leaves, Aust. J. Plant. Physiol., 9, 1, 121–137, 1982.
Feng, X. and Epstein, S.: Carbon isotopes of trees from arid environments and implications for reconstructing atmosphere CO_2 concentration, Geochim. Cosmochim. Ac., 59, 12, 2599–2608, 1995.

Feng, X., Peterson, J. C., Quideau, S. A., Virginia, R. A., Graham, R. C., Sonder, L. J., and Chadwick, O. A.: Distribution, accumulation and fluxes of soil carbon in four monoculture lysimeters at San Dimas Experimental Forest, California, Geochim. Cosmochim. Ac., 63, 9, 1319–1333, 1999.

Feng, X.: A theoretical analysis of carbon isotope evolution of decomposing plant litters and soil organic matter, Global Biogeochem. Cy., 16, 4, 1119, doi:10.1029/2002GB001867, 2002.

Fernandez, I., Mahieu, N., and Cadisch, G.:. Carbon isotopic fractionation during decomposition of plant materials of different quality, Global Biogeochem. Cy., 17, 1075, doi:10.1029/2001GB001834, 2003.

Frakes, L. A. and Sun, J. Z.: A carbon isotope record of the upper Chinese loess sequence: estimates of plant types during stadials and interstadials, Palaeogeogr. Palaeocl., 108, 2, 183–189, 1994.

Friedli, H., Moore, E., Oeschger, H., Siegenthaler, U., and Stauffer, B.: $^{13}C/^{12}C$ ratios in CO_2 extracted from Antarctic ice, Geophys. Res. Lett., 11, 5, 1145–1148, 1984.

Ganopolski, A., Rahmstorf, S., Petoukhov, V., and Claussen, M.: Simulation of mordern and glacial climates with a coupled global model of intermediate complexity, Nature, 391, 6665, 351–356, 1998.

Ghannoum, O., von Gaemmerer, S., and Conory, J. P.: The effect of drought on plant water use efficiency of nine NAD–ME and nine NADP–ME Australian C_4 grasses, Funct. Plant Biol., 29, 7, 1337–1348, 2002.

Gregorich, E. G., Ellert, B. H., and Monreal, C. M.: Turnover of soil organic matter and storage of corn residue carbon estimated from natural ^{13}C abundance, Can. J. Soil Sci., 75, 2, 161–167, 1995.

Gu, Z.: The carbonate isotopic composition of the loess – paleosol sequence and its implication of paleoclimatic change, Chinese Sciences Bulletin, 36, 10, 1979–1983, 1991.

Guillaume, K., Huard, M., Gignoux, J., Mariotti, A., and Abbadie, L.: Does the timing of litter inputs determine natural abundance of ^{13}C in soil organic matter? Insights from an African tiger bush ecosystem, Oecologia, 127, 295–304, 2001.

Han, J. M., Keppens, E., Liu, T. S., Paepe, R., and Jiang, W. Y.: Stable isotope composition of the carbonate concretion in loess and climate change, Quatern. Int., 37, 1, 37–43, 1996.

Henderson, S. A., von Caemmerer, S., and Farquhar, G. D.: Short-term measurements of carbon isotope discrimination in several C_4 species, Aust. J. Plant. Physiol., 19, 263–285, 1992.

Institute of Botany: CAS Flora of China (in Chinese), Chinese Science Press, Beijing, 1982.
Krull, E. S., Bestland, E. A., and Gates, W. P.: Soil organic matter decomposition and turnover in a Tropical Ultisol: Evidence from δ^{13}C, δ^{15}N and Geochemistry, Radiocarbon, 44, 1, 93–112, 2005.
Körner, C., Farquhar, G., and Dand Roksandic, Z.: A global survey of carbon isotope discrimination in plants from high altitude, Oecologia, 74, 6, 623–632, 1988.
Leuenberger, M., Siegenthaler, U., and Langway, C. C.: Carbon isotope composition of atmospheric CO_2 during the last ice age from an Antarctic ice core, Nature, 357, 6385, 488–490, 1992.
Lin, B., Liu, R., and An, Z.: Preliminary research on stable isotopic composition of Chinese loess, in: Loess, Environment and Global Chang, edited by: Liu, T., Science Press, Beijing, 124–131, 1991.
Liu, M.: Composition of carbon isotope of plants in Donglingshan Mountain, North China, MS. thesis, Peking University, 2003.
Liu, W. G., Feng, X. H., Ning, Y. F., Zhang, Q. L., and Cao, Y. N.: δ^{13}C variation of C_3 and C_4 plants across an Asian monsoon rainfall gradient in arid northwestern China, Global Change Biol., 11, 1094–1100, doi:10.1111/j.1365-246.2005.00969.x., 2005.
Macko, A. and Estep, M. L. F.: Microbial alteration of stable nitrogen and carbon isotopic compositions of organic matter, Org. Geochem., 6, 787–790, 1984.
Marino, B. D., McElroy, M. B., Salawitch, R. J., and Spaulding, W. G.: Glocial to interglacial variations in the carbon isotopic composition of atmospheric CO_2, Nature, 357, 6385, 461–466, 1992.
Natelhoffer, K. J. and Fry, B.: Controls on natural nitrogen-15 and carbon-13 abundances in forest soil organic matter, Soil Sci. Soc. Am. J., 52, 6, 1633–1640, 1988.
Poage, M. A. and Feng, X.: A theoretical model of steady state δ^{13}C vs. depth profiles of decomposing soil organic matter, Global Biogeochem. Cy., 18, GB2016, doi:10.1029/2003GB002195, 2004.
Polley, H. W., Johnson, H. B., Marino, B. D., and Mayeux, H. S.: Increase in C_3 plant water-use efficiency and biomass over Glacial to present CO_2 concentrations, Nature, 361, 6433, 61–64, 1993.
Schleser, G. H. and Pohling, R.: δ^{13}C record in forest soil using a rapid method for preparing carbon dioxide samples, Int. J. Appl. Radiat. Is., 31, 7, 769–773, 1980.
Schulze, E.-D., Ellis, R., Schulze, W., and Ziegler, H.: Diversity, metabolic types and δ^{13}C

carbon isotope ratios in the grass flora of Namibia in relation to growth form, precipitation and habitat conditions, Oecologia, 106, 352–369, 1996.
Schulze, E.-D., Turner, N. C., Nicolle, D., and Schumacher, J.: Leaf and wood carbon isotope ratios, specific leaf areas and wood growth of Eucalyptus species across a rainfall gradient in Australia, Tree Physiol., 26, 5, 479–492, 2006.
Schwartz, D., Mariotti, R., Lanfranchi, R., and Guillet B.: $^{13}C/^{12}C$ ratios of soil organic matter as indicators of vegetation changes in the Congo, Geoderma, 39, 2, 97–103, 1986.
Shen, C. D., Yi, W. X., Sun, Y. M., Xing, C. P., Yang, Y., Peng, S. L., and Li, Z. A.: ^{14}C apparent ages and $\delta^{13}C$ distribution of forest soils in Dinghushan Natural Reserve, Quatern. Sci., 20, 4, 335–344, 2000 (in Chinese with English abstract).
Smedley, M. P., Dawson, T. E., Comstock, J. P., Donovan, L. A., Sherrill, D. E., Cook, C. S., and Ehleringer, J. R.: Seasonal carbon isotope discrimination in a grassland community, Oecologia, 85, 314–320, 1991.
Stanley, H. A. and Nancy, E. S.: Soil carbon isotope evidence for Holocene habitat change in the Kenya Rift Valley, Science, 253, 1402–1405, 1991.
Stewart, G. R., Turnbull, M. H., Schmidt, S., and Erskine, P. D.: ^{13}C Natural abundance in plant communities along a rainfall gradient: a biological integrator of water availability, Aust. J. Plant. Physiol., 22, 1, 51–55, 1995.
Stout, J. D. and Rafter, T. A.: The $^{13}C/^{12}C$ isotopic ratios of some New Zealand tussock grassland soils, in: Stable Isotopes in the Earth Sciences, edited by: Robinson, B. W., DSIR BULL, Wellington, 75–83, 1978.
Stout, J. D., Rafter, T. A., and Throughton, J. H.: The possible significance of isotopic ratios in paleoecology, in: Quaternary Studies, edited by: Suggate, R. P., Cresswell, M., Royal Society of New Zealand, Wellington, 279–286, 1975.
Torn, M. S., Lapenis, A. G., Timofeev, A., Fischer, M., Babikov, B., and Harden, J. W.: Organic carbon and carbon isotopes in modern and 100-year-old-soil archives of the Russian steppe, Glob. Change Biol., 8, 10, 941–953, 2002.
Trolier, M., White, J. W., Tans, P. P., Masarie, K. A., and Gemery, P. A.: Monitoring the isotopic composition of atmospheric CO_2: measurements from the NOAA Global Air Sampling Network, J. Geophys. Res., 100, 10, 25 897–25 916, 1996.
Troughton, J. H., Stout, J. D., and Rafter, T. A.: Long-term stability of plant communities, Carnegie Inst Wash Yearbook, 73, 838–845, 1974.
Vidic, N. J. and Montaez, I. P.: Climatically driven glacial-interglacial variations in C_3 and C_4

plant proportions on the Chinese Loess Plateau, Geology, 32, 4, 337–340, 2004.
Wang, G. A., Han, J. M., and Liu, T. S.: The carbon isotopic composition of C_3herbaceous plants in loess area of North China, Science in China in Series D- Earth Sciences, 46, 10, 1069–1076, 2003.
Wang, G. A., Han, J. M., Zhou, L. P., Xiong, X. G., and Wu, Z. H.: Carbon isotope ratios of plants and occurrences of C_4 species under different soil moisture regimes in arid region of Northwest China, Physiol. Plantarum, 125, 1, 74–81, 2005.
Wang, G. A., Han, J. M., Zhou, L. P., Xiong, X. G., Tan, M., Wu, Z. H., and Peng, J.: Carbon isotope ratios of C_4 plants in loess areas of North China, Science in China in Series D- Earth Sciences, 49, 1, 97–102, 2006.
Wang, G. A.: δ^{13}C values in herbaceous plants and surface soil organic matter in north China. Ph.D. thesis in the Institute of Geology and Geophysics, Chinese Academy of Sciences, 2001.
Wang, H. and Follmer, L. R.: Proxy of monsoon seasonality in carbon isotopes from paleaosols of the southern Chinese Loess Plateau, Geology, 26, 11, 987–990, 1998.
Wang, Y. and Zheng, S. H.: Paleosol nodules as Pleistocene paleoclimatic indicators, Luochuan, P. R. China, Palaeogeogr. Palaeocl., 76, 1, 39–44, 1989.
Wedin, D. A, Tieszen, L. L., Dewey, B., and Pastor, J.: Carbon isotope dynamics during grass decomposition and soil organic matter formation, Ecology, 76, 5, 1383–1392, 1995.
Wu, N. Q., Lu, H. Y., and Sun, X. J.: Climatic factor transfer function from opal phytolith and its application in paleoclimate reconstruction of China loess- palaeosol sequence, Scientia Geologia Sinica, 1(Supplement), 105–114, 1995.
Zhang, Z. H., Zhao, M. X., Yang, X. D., Wang, S. M., Jiang, X. Z., Oldfield, F., and Eglinton, G.: A hydrocarbon biomarker record for the last 40 Kyr of plant input to Lake Heqing, Southwestern China, Org. Geochem., 35, 5, 595–613, 2004.
Zinke, P. J., Stangenberger, A. G., Post, W. M., Emanuel, W. R., and Olson, J. S.: Worldwide Organic Soil Carbon and Nitrogen Data. NDP-018 http://cdiac.esd.ornl.gov/ndps/ndp018.html. Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, Tennessee, USA, 1986.
Zinke, P. J., Stangenberger, A. G., Post, W. M., Emanuel, W. R., and Olson, J. S.: Global Organic Soil Carbon and Nitrogen (Zinke et al.). Data set. Available on-line http://www.daac.ornl.gov from Oak Ridge National Laboratory Distributed Active Archive Center, Oak Ridge, Tennessee, USA, 1998.

Table 1. Observed carbon isotopic differences between vegetation or plant litter and soil organic matter.

	Site 1		Site 2		Site 3	
	$\delta^{13}C_{sur-veg}$	$\delta^{13}C_{botm-veg}$	$\delta^{13}C_{sur-litter}$	$\delta^{13}C_{botm-litter}$	$\delta^{13}C_{sur-veg}$	$\delta^{13}C_{botm-veg}$
Profile 1	0.7	5.2	1.5	3.7	2.5	3.9
Profile 2	1.0	3.7	2.1	3.8	1.4	3.4
Profile 3	0.6	5.9	2.4	3.7	2.2	3.3
Profile 4	1.0	5.2	3.0	5.0		
Profile 5	1.1	2.6	2.5	4.5		
Profile 6	1.0	3.2				
Mean	0.9	4.3	2.3	4.1	2.0	3.5
Stand. Error	0.1	0.5	0.2	0.2	0.3	0.2

The subscripts: sur=surface soil; botm=bottom soil; veg=vegetation; litter=plant litter at the soil surface.

Table 2. Comparison of C_4% contribution to paleovegetation with various corrections.

Holocene			Using Eq. (1)*	Using Eq. (3)*
	$\delta^{13}C_{C_3}$	$\delta^{13}C_{C_4}$	C_4%	C_4%
Today's vegetation	−27.5	−12.5	53.3	41
$\delta^{13}C_{atm}$ correction (+1.3)	−26.2	−11.2	44.7	32.7
CO_2 concentration correction (C_3: +1.6‰)	−24.6	−11.2	38.1	24.6
Last Glaciation			Using Eq. (1)*	Using Eq. (3)*
	$\delta^{13}C_{C_3}$	$\delta^{13}C_{C_3}$	C_4%	C_4%
Today's vegetation	−27.5	−12.5	34	22
$\delta^{13}C_{atm}$ correction (+1.3‰–0.3‰)	−26.5	−11.5	27.3	15.3
Precipitation correction (C_3: +0.8‰)	−25.7	−11.5	23.2	10.6
CO_2 concentration correction (C_3: +1.6‰+1.6‰)	−22.5	−11.5	0.9	−15.5

* Calculations assume that $\delta^{13}C_{SOM}$ is −19.5‰ for Holocene and −22.4‰ for LG.

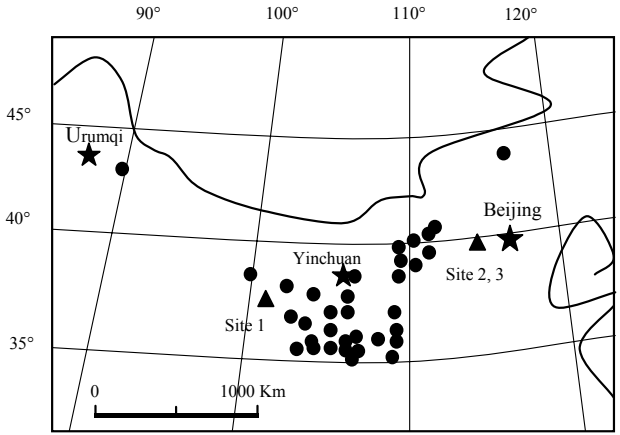

Fig. 1. Sample sites of this study. Solid circles and triangles represent plant and soil profile sites, respectively.

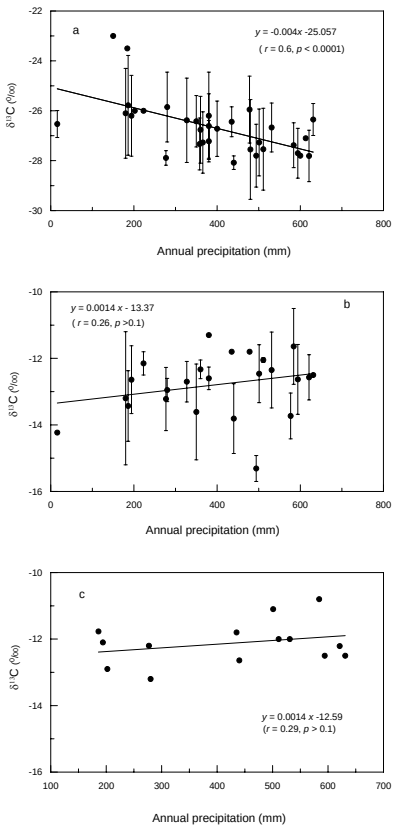

Fig. 2. δ^{13}C variations of plant leaves with the annual precipitation amount along a rainfall gradient in North China. **(a)** Variations of site-averaged δ^{13}C in C_3 plants. **(b)** Variations of site-averaged δ^{13}C in C_4 plants. **(c)** δ^{13}C variations of *Setaria viridis* (a C_4 species). Each bar represents the mean δ^{13}C value averaged for a given site and the standard deviation of that site.

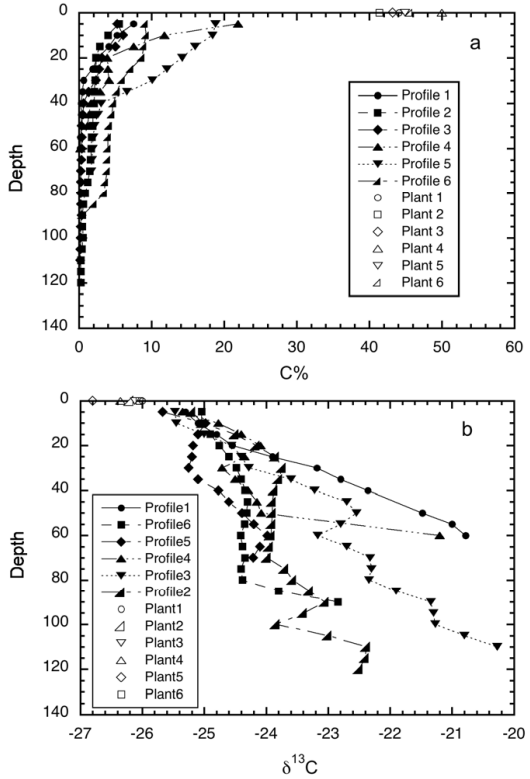

Fig. 3. Carbon contents of soil and aboveground plants **(a)** and δ^{13}C values of soil organic matter and aboveground plants **(b)** at Site 1. The organic carbon contents and δ^{13}C values at 0 cm depth of each profiles are those for aboveground plants within each plots.

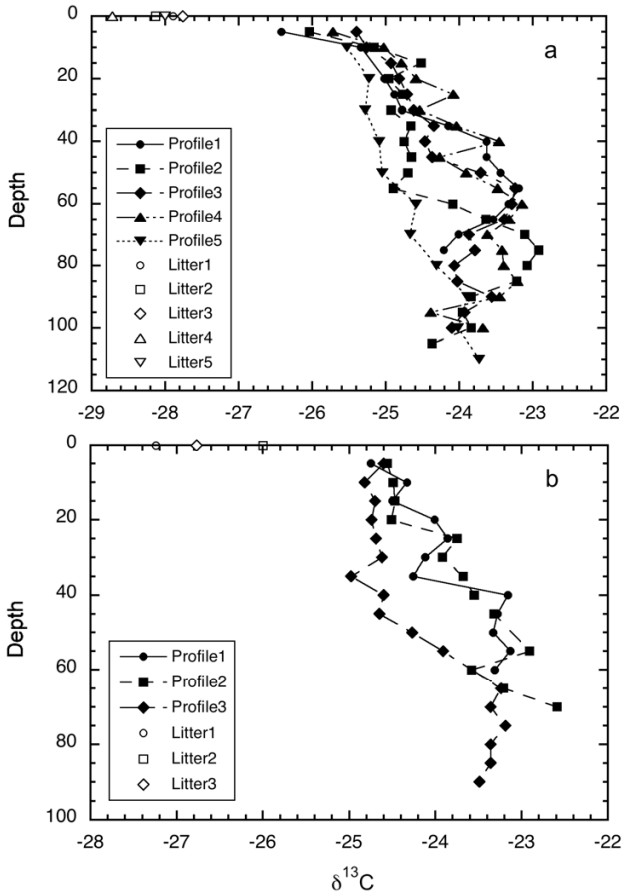

Fig. 4. Variations of δ^{13}C values of organic matter with depths at **(a)** Site 2, and **(b)** Site 3. The δ^{13}C value at the 0 cm depth of each profile is that of litter within each plot in (a), and aboveground plants in (b).