High-resolution summer precipitation variations in the western Chinese Loess Plateau during the last glacial

Zhiguo Rao1, Fahu Chen1, Hai Cheng2, Weiguo Liu3, Guo’an Wang4, Zhongping Lai5 & Jan Bloemendaal6

We present a summer precipitation reconstruction for the last glacial (LG) on the western edge of the Chinese Loess Plateau (CLP) using a well-dated organic carbon isotopic dataset together with an independent modern process study results. Our results demonstrate that summer precipitation variations in the CLP during the LG were broadly correlated to the intensity of the Asian summer monsoon (ASM) as recorded by stalagmite oxygen isotopes from southern China. During the last deglaciation, the onset of the increase in temperatures at high latitudes in the Northern Hemisphere and decline in the intensity of the East Asia winter monsoon in mid latitudes was earlier than the increase in ASM intensity and our reconstructed summer precipitation in the western CLP. Quantitative reconstruction of a single paleoclimatic factor provides new insights and opportunities for further understanding of the paleoclimatic variations in monsoonal East Asia and their relation to the global climatic system.

Loess/paleosol sequences in the Chinese Loess Plateau (CLP), especially those with high accumulation rates in the northern and western CLP, are among the best terrestrial archives for late Quaternary paleoclimatic studies1–2. Paleoclimatic records in this area are extremely important for the understanding of the linkage and interplay between the evolution of the ASM at low latitudes3 and temperature variations at high latitudes4. However, until now, a high resolution paleoprecipitation reconstruction using a clear driving mechanism has been lacking in the western CLP.

Fractionation of carbon isotopes by modern C3 plants during CO2 uptake and fixation (Δ) can be described by mathematical models using the formula Δ = a + (b – a)(P1/P2), where a is the carbon isotopic fractionation during CO2 diffusion (ca. 4.4%), b is the net fractionation caused by carboxylation (ca. 29%) and P1 and P2 are the intercellular and ambient partial pressures of CO2, respectively5. In principle, increasing precipitation will result in more negative carbon isotopic composition (δ13C) values of C3 plants, due to the relatively high stomatal conductance present under a humid environment which will increase P1 and Δ, and vice versa.

The relations between the δ13C of modern C3 plants and environmental factors have been widely studied and consistent results have been obtained which indicate that the δ13C of modern C3 plants responds principally to precipitation variations6–8 (Supplementary Figs. S1, S2 and S3). To a lesser extent, it also reflects more minor changes in temperature, altitude, and latitude9,10. Clearly, if the local terrestrial ecosystem was dominated by, or composed entirely of, C3 plants during a specific interval, the sedimentary δ13C data that derives from the local terrestrial biomass should be an indicator of precipitation. In Western Europe, total organic carbon isotopic data (δ13C(OTOC)) of loess indicate a great predominance of C3 plants since the LG, and therefore these δ13C(OTOC) data have been used for precipitation reconstruction11,12.

During the past decade, the δ13C(OTOC) record of more than 10 Chinese loess/paleosol profiles has been studied (Fig. 1). The results demonstrate that, along the temporal sequence, the relative abundance of C4 plants increased from the LG to the Holocene (Supplementary Fig. S4); and that, along a spatial gradient, relative C4 plant abundance decreased gradually northwards (Supplementary Fig. S5)13–17, as is the case today18. During the LG, δ13C(OTOC) data from the high-temperature eastern CLP (to the east of the Liupan Mts. Fig. 1) indicate that this
area was dominated by C₃ plants with a minor increase in C₄ plants during marine isotope stage 3 (Supplementary Fig. S6), which means that the vegetation in the western CLP during the LG could be expected to be predominantly or entirely composed of C₃ plants (Supplementary Fig. S6).

The Yuanbao profile (YB, 103.6°E, 35.15°N, 2,040 m a.s.l.), close to the Tibetan Plateau, is located on the western edge of the CLP (Fig. 1) with modern mean annual temperature (MAT) and precipitation (MAP) of ca. 6.8°C and ca. 500 mm, respectively. The Jingyuan profile (JY, 104.6°N, 36.35°N, 2,210 m a.s.l.), close to the Tengger Desert, is located on the northwestern edge of the CLP (Fig. 1) with modern MAP and MAT of ca. 238 mm and ca. 5.2°C, respectively. At both sites, precipitation primarily occurs during May to September (ca. 80% of MAP) with higher temperatures (also the main growing season of the local terrestrial vegetation), consistent with the modern climate in this area as controlled principally by the Asian monsoon (Supplementary Fig. S7 and S8). High resolution optically stimulated luminescence (OSL) dating clearly demonstrates that the loess/paleosol sequences of YB and JY have accumulated continuously since the LG without significant hiatuses. Previous studies have demonstrated that the loess/paleosol sequences located in the western CLP, including YB and JY, have great potential for high-resolution paleoclimatic reconstruction during the LG. The top 25.74 m of YB was sampled for δ¹³C_TOC study at a 4 cm interval (ca. 100 years per sample) and the top ca. 30 m of JY was sampled at a 40 cm interval (ca. 1000 years per sample) for δ¹³C_TOC study. High resolution δ¹³C_TOC data (Fig. 2) from YB indicated that the study site was dominated by C₃ plants during the LG and the relative abundance of C₄ plants slightly increased during the Holocene period (Supplementary Fig. S6). δ¹³C_TOC data from JY (Fig. 2) indicated that the vegetation in the study site was dominated by C₃ plants during both the LG and the Holocene (Supplementary Fig. S6). These results are consistent with the previous finding that the relative abundance of C₄ plants decreased northwestwards across the entire CLP (Supplementary Fig. S4, S5). All of this evidence demonstrates that the δ¹³C_TOC data from YB during the LG and those from JY since the LG are suitable for paleoprecipitation reconstruction and that the results from both sites can be directly compared. For paleoprecipitation reconstruction, the quantitative relations between δ¹³C of modern C₃ plants and precipitation are established directly from modern case studies (Fig. 3). However, although almost all of
the relevant studies have shown a negative correlation between the 
$\delta^{13}$C of C$_3$ plants and precipitation, the magnitude of the correlation
varies between different C$_3$ plant species and in different regions$^6$–$^9$
(Supplementary Fig. S2, S3 and Table S1). Here we have used the
results of $\delta^{13}$C$_{TOC}$ measurements of 196 modern surface soils from
arid central Asia$^{27,28}$ as our point of reference (Supplementary Fig.
S9). Our selection is based on: (i) the close location of these surface
soils to the JY and YB sites; and (ii) the fact that both investigation on
modern plants$^{29}$ and the $\delta^{13}$C$_{TOC}$ data of these surface soils$^{27,28}$ (most
of them, $\geq 24\%$) demonstrate that the modern terrestrial ecosys-
tem in this area is greatly dominated by C$_3$ plants (Supplementary
Fig. S9). Considering that the $\delta^{13}$C of modern C$_3$ plants in arid central
Asia principally responds to summer precipitation and the con-
strained uncertainties of the linear correlation between averaged
surface soil $\delta^{13}$C$_{TOC}$ values of 19 weather stations and the summer
precipitation amount (May to Sept.) as recorded by the correspond-
ing weather stations, we chose the correspondingly quantitative
relation$^{27}$ for the reconstruction of summer precipitation (Supple-
mentary Fig. S9). Because the carbon isotopic composition of organic
matter becomes more positive after long-term decomposition$^{30}$, we
have selected 1% as the magnitude of the baseline value for the
$\delta^{13}$C$_{TOC}$ data of loess/paleosol samples relative to their original
values at the time of deposition. The secondary effect of temperature
and atmospheric CO$_2$ concentrations on C$_3$ plant $\delta^{13}$C has been
neglected. Based on these assumptions our final expression for sum-
mer precipitation reconstruction is: summer precipitation (mm)
$= 58*(\delta^{13}$C$_{TOC}$*1000 + 1) - 1266.5.

Correspondingly, the 95% confidence interval of the relevant linear
relation$^{27}$ is used for the uncertainty estimation of the summer
precipitation reconstruction (Supplementary Fig. S9).

**Results**

In spite of the different resolutions, a high variability occurs in sum-
mer precipitation reconstructions of both the YB and JY sites, appar-
ently consistent with the characteristics of modern climate in the
study area (Supplementary Fig. S10); this indicates the high sensitiv-
ity of carbon isotopes of local terrestrial vegetation to variations in
summer precipitation.

Our results indicate that summer precipitation varied from ca.
20 mm to 150 mm with uncertainties of ca. 70 $\sim$ 80 mm at the JY
site during the past 70 ka with most values during the LG of less than
100 mm. Summer precipitation decreased gradually from MIS3 to
an extremely arid MIS2, and then increased towards the Holocene (Supplementary Fig. S11). Considering the location of the JY profile so close to the Tengger Desert (Fig. 1), an extremely arid climate at the JY site during MIS2 is reasonable. Although there are only a few data values for the Holocene, the results indicate that the maximum summer precipitation during the Holocene was more than 150 mm (with uncertainty of ca. 70 mm), which is close to the modern summer precipitation in the JY area (ca. 190 mm, averaged from 1961 to 1990) (Supplementary Fig. S10).

Reconstructed summer precipitation at the YB site during the LG varied mainly from ca. 150 mm to 350 mm with uncertainties of ca. 70 mm (Supplementary Fig. S11, Table S2). Modern summer precipitation in the YB area is ca. 400 mm (averaged from 1961 to 1990, Supplementary Table S3), greater than that of the JY site. During the LG, reconstructed summer precipitation at YB was generally higher than that at JY (Supplementary Fig. S10), suggesting that the modern climatic gradient between the JY and YB sites is consistent with that during the LG, which supports our methodology.

Loess δ13CTOC data of the YB profile during the Holocene are apparently contaminated by C4 plants, especially during the early Holocene (Fig. 2, Supplementary Fig. S6), and this is the essential reason why a reconstruction of summer precipitation for the whole Holocene at the YB site has not been completed. However, if loess δ13CTOC data of the topmost 2 samples (−28% and −27.5% respectively, Supplementary Table S2) of the YB profile were to be used for the calculation of summer precipitation, then summer precipitation levels of ca. 415 mm and 385 mm (with uncertainties of ca. 70 mm) respectively would be obtained, consistent with modern summer
precipitation at the YB site of an averaged value of ca. 400 mm; these findings also support our methodology.

A series of rapid climatic events during the LG has been recorded at the YB site using a grain size proxy for winter monsoon intensity. These events may be associated with temperature variations at high latitudes of the Northern Hemisphere and a relation between the intensity of the winter monsoon and the north westerlies. Based on the corresponding North Greenland Ice Core Project (NGRIP) ice-core ages (GICC05) of Heinrich and interstadial events, we adjusted the ages of the YB winter monsoon record based on their correlation with NGRIP oxygen isotopic (δ18O) record (Supplementary Fig. S12). There is an overall consistency between the OSL dating results of the YB profile and the transferred NGRIP ice-core ages, with relatively younger (ca. 4 ~ 5 ka) OSL ages occurring between 20 ka to 60 ka (Supplementary Fig. S13).

Summer precipitation along the adjusted NGRIP ice-core age series at the YB site shows a high degree of consistency with the stalagmite δ18O record from Hulu Cave in southern China. Both data sets record Heinrich events 1 to 6 and interstadial events 1 to 19 (Fig. 3); in general, arid events, represented by sharply decreased summer precipitation in the western CLP, broadly correspond to a weakened ASM recorded in stalagmite δ18O data from southeastern China, and to cold events in high northern latitudes recorded in the NGRIP δ18O data. More importantly, from the H4 to H1 events, the stalagmite δ18O data gradually became more positive and the summer precipitation in the YB area also decreased gradually (Fig. 3). Assuming that the stalagmite δ18O data from southern China have recorded variations in the intensity of the ASM, and based on the comparison of the YB and Hulu records, it appears that the decrease in the intensity of the ASM was the direct cause of the decrease in summer precipitation in northwest China.

**Discussion**

The assumption that the Chinese stalagmite δ18O data record the intensity of the ASM has been widely debated. For example, in a recent comparative simulation study of the Last Glacial Maximum (LGM) and Heinrich event 1 (H1), changes in the Indian monsoon that are controlled by the sea surface temperature (SST) of the Indian Ocean have been suggested as the main cause of the stalagmite δ18O variations in China. Due to the lack of direct evidence that the Indian Ocean SST during H1 was lower than during the LGM, and based on the high degree of consistency between the variations in our reconstructed summer precipitation and those in the southern Chinese stalagmite δ18O data (Fig. 3), our results demonstrate that, at least during the LG period, variations in the Chinese stalagmite δ18O data records are a valid indicator of the intensity of the ASM.

During the transition from the LG to the Holocene, summer precipitation in the YB area and the intensity of the ASM decreased gradually. Subsequently, an enhancement of both the ASM and summer precipitation occurred around the time of the H1 event (Fig. 3). The onset of the rise of temperature at high latitudes recorded by the NGRIP δ18O data occurred at around the time of the H2 event or earlier (Fig. 3). Grain size data from the YB profile indicate that the onset of the decrease of the East Asian winter monsoon was also coeval with the H2 event (Fig. 3), which is also consistent with grain size records from the Luochuan and JY2 profiles. During the transition from the LG to the Holocene, it seems that both high latitude temperatures and the mid latitude East Asian winter monsoon responded rapidly to increasing insolation (Fig. 3). However, the response of the low latitude ASM was apparently delayed. All of this evidence indicates more complex driving mechanisms for the low latitude summer monsoon than for the mid and high latitude climatic components.

In conclusion, our results are significant in the following respects: 1) because previous studies have already demonstrated that the local biomass on the western edge of the CLP and in the loess area in Western Europe was dominated by C3 plants since the LG, sedimentary δ13C derived therefrom may be a valid indicator of paleoprecipitation; and therefore our work provides a new paradigm for paleoprecipitation reconstruction in this vast region between Western Europe and the western CLP and the regions to their north; 2) the paleoclimatic significance of the Chinese stalagmite δ18O data has been widely debated; however, we demonstrated that at least during the LG, the Chinese stalagmite δ18O data are a credible indicator of the intensity of the ASM. Further work needs to be done to test the validity of this finding over longer timescales; and 3) the combination of future independent paleotemperature reconstructions and our paleoprecipitation reconstruction results may constitute a major advance in paleoclimatic studies of monsoonal East Asia.

**Methods**

The YB profile was sampled at 2 cm intervals (ca. 50 years per sample) for the top 25.74 m. After removal of organic matter and carbonate with HCl (−10%) and H2O2 (−10%), the grain size of all the samples was measured using a Malvern Master Sizer 2000 laser diffraction analyser. The measurement range of this equipment is 0.02–2000 μm. In this paper we use the percentage of the grain size fraction > 40 μm as an indicator of the east Asian winter monsoon intensity and the strength of the north westerlies. A 4 cm interval was used for the δ18O measurements. The samples were pretreated as follows: ~ 10% hydrochloric acid (HCl) was used to remove carbonates, followed by washing with distilled water until the suspension was neutral. The wet samples were then sieved at 120 μm to remove tiny sand particles and gravels. After sieving, the samples were dried at 70 °C. The gas collection method involved static combustion. The evolved CO2 was analyzed for δ18O using a Thermo Finnigan Delta Plus mass spectrometer. The standard materials used for the measurements were international standard tree-rings (Corundum balls, IAEA-CS5). The mean value of 29 repeat measurements was −25.7% ± 0.3% (the reported value of the standard is −25.49 ± 0.72‰). Repeated measurements on both the standard materials and samples (81 times in total) showed that the experimental error is less than ±0.2%. (See references 19 and 20 for the OSL dating methods used on profile YB, and references 21 and 22 for those used on profile JY. See reference 25 for the organic carbon isotopic data for profile JY.)
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**Author contributions**

F.H.C. designed the study and led the writing of the paper. Z.G.R. performed the organic carbon isotopic analysis of the JY profile. W.G.L. performed the organic carbon isotopic analysis of the YB profile. All authors contributed to the discussion and interpretation of the results and to the writing of the manuscript.

**Additional information**

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