$\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records from a Cenozoic sedimentary sequence in the Lanzhou Basin, Northwestern China: Implications for palaeoenvironmental and palaeoecological changes

Baofeng Li* Donghuai Sun Xin Wang Yuebao Zhang Wenwei Hu

Fei Wang Zaijun Li Zhiwei Ma Baiqing Liang

Key Laboratory of West China's Environmental System (Ministry of Education), Lanzhou

University, Lanzhou 730000, China

*Corresponding author: E-mail: *libf11@lzu.edu.cn*

Present address: 222 South Tianshui Road, Lanzhou, Gansu, China

Abstract

In northwestern China, carbonate $\delta^{18}\text{O}$ variation has been closely associated with evaporation and precipitation, whereas the variation of carbonate $\delta^{13}\text{C}$ generally reflects patterns of palaeovegetation. Located within the transitional zone between the Chinese Loess Plateau and the Tibetan Plateau, the Lanzhou Basin has developed a continuous sequence of Cenozoic sediments which have been subjected to detailed sedimentological and high-resolution magnetostratigraphic analyses. In the present study, pedogenic carbonate O and C isotopic analyses were obtained throughout the entire Cenozoic sequence. The $\delta^{18}\text{O}$ record exhibits a general positive trend with several abrupt changes. A dramatic positive shift in the $\delta^{18}\text{O}$ record at ~33 Ma indicates the initiation of the aridification process within the basin, which was likely associated with the late Eocene westward retreat of the Tethys Sea and global cooling. Two significant positive shifts in the $\delta^{18}\text{O}$ record at ~22 Ma and ~3.5 Ma are synchronous with major increases in aeolian dust deposition on the Chinese Loess Plateau and in the North Pacific Ocean, suggesting the intensified aridity of the Asian interior, which is likely related to the stepwise uplift of the Tibetan Plateau via the blocking of water vapour pathways. The $\delta^{13}\text{C}$ values exhibit a weak positive trend with a remarkable shift at ~3.5 Ma. This trend is likely related to a decrease in vegetation density in response to the ongoing Cenozoic aridification, whereas the shift at ~3.5 Ma may reflect the large-scale expansion of C_4 plants.

Keywords: Aridification; Palaeoecosystem; Pedogenic carbonate; Oxygen and carbon

isotopes

1. Introduction

The uplift of the Tibetan Plateau, the retreat of the Tethys Sea and ongoing global cooling are considered the three major causes of aridification in the Asian interior during the Cenozoic due to their effects on water vapour transport in Asia (e.g. Dupont-Nivet et al., 2007; Lu et al., 2010; Xiao et al., 2010; Abels et al., 2011; Licht et al., 2014). The nature of the aridification process has been revealed using the downwind records of aeolian deposits from the Chinese Loess Plateau (CLP) (e.g. Sun et al., 1998a, 1998b; Ding et al., 1999; Guo et al., 2002; Qiang et al., 2011) and the North Pacific Ocean (e.g. Rea et al., 1985, 1998). Various analytical techniques have been used to investigate these sedimentary archives, including measurements of grain-size, magnetic properties and elemental composition (e.g. Gallet et al., 1996; Sun D. et al., 2004; Ding et al., 2005). In comparison with these conventional approaches, carbonate $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values have been rarely used, despite their utility as a palaeoenvironmental proxy (Cerling, 1984; Cerling and Quade, 1993; Quade and Cerling, 1995; Liu et al., 1996).

Systematic studies of aeolian sequences from the central CLP suggested that deposition of aeolian sediments commenced at ~7-8 Ma (Sun et al., 1998a, 1998b; Ding et al., 1999; Fig. 1). Studies on the western CLP extended the Chinese aeolian dust history back to 26-22 Ma (Guo et al., 2002; Qiang et al., 2011; Zhang et al., 2014; Fig. 1), indicating that the aridification of the Asian interior initiated in the late Oligocene. Chronological and conventional proxy index studies of the 3200-m-thick Cenozoic sequence in the Lanzhou Basin have revealed that significant intensification

of the regional aridification process occurred at ~26, ~22, ~14, ~7.2 and ~2.6 Ma (Sun et al., 2011b; Zhang et al., 2014). In contrast to conventional proxy indices, the carbonate $\delta^{18}\text{O}$ record can provide more detailed information on regional temperature, precipitation and evaporation (e.g. Cerling, 1984; Cerling and Quade, 1993; Quade et al., 1995; Liu et al., 1996; Hsieh et al., 1998), especially in arid areas and in strata with multiple sedimentary facies. In addition, the carbonate $\delta^{13}\text{C}$ record can be used to indicate the nature of local vegetation in terms of the C_3/C_4 biomass ratios (e.g. Cerling, 1984; Quade and Cerling, 1995).

(Fig. 1)

Cerling et al. (1988) used pedogenic carbonate $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ records in the Turkana Basin to reconstruct the local ecological variability and regional aridity history. Quade et al. (1989a) reconstructed the evolutionary history of the Asian monsoon system and the corresponding palaeoecology since 18 Ma using pedogenic carbonate $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ records from Pakistan. Following this pioneering work, a series of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ records have been used to reconstruct local and/or regional palaeoecologies and palaeoenvironments (e.g. Smith et al., 1993; Koch et al., 1995; Levin et al., 2011; Singh et al., 2012; Caves et al., 2014). A study of pedogenic carbonate $\delta^{13}\text{C}$ records from the CLP revealed ages associated with the expansion of C_4 plants ranging from the Pliocene to the Pleistocene (e.g. Ding and Yang, 2000; Jiang et al., 2002; An et al., 2005; Kaakinen et al., 2006; Sun et al., 2012). A large carbonate $\delta^{18}\text{O}$ data set from the northern margin of the Tibetan Plateau revealed a “Neogene positive trend”, which was interpreted as a signal of regional aridification

(Kent-Corson et al., 2009). Similarly, many other pedogenic carbonate $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ records from sedimentary basins surrounding the Tibetan Plateau have been used to explore regional palaeoclimates and palaeoecosystem evolutions from different perspectives and on different time scales (e.g. Dettman et al., 2003; Graham et al., 2005; Wang and Deng, 2005; Hough et al., 2011; Zhuang et al., 2011; Sun et al., 2013; Han et al., 2014).

In this paper, pedogenic carbonate $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records from the Lanzhou Basin, Northwestern China, are presented to better understand the history of regional aridity and ecosystem development, as well as the driving mechanisms of these palaeoenvironmental and palaeoecological changes.

2. Cenozoic sedimentary sequence and chronology of the Lanzhou Basin

The Lanzhou Basin is located in a transitional zone between the Tibetan Plateau and the CLP (Fig. 1). The mean annual temperature (MAT) is 8-10 °C, and the mean annual precipitation (MAP) is less than 300 mm, with more than 70% falling during the summer months. The mean evaporation is approximately 1500 mm.

The Cenozoic sedimentary sequence unconformably overlies the Cretaceous sandstone. It is divided into the Xiliugou, Yehucheng, Xianshuihe and Linxia Formations, and Quaternary loess sediments (Fig. 2). The interpretation of the depositional environment and the age constraints used in this study are based on Sun et al. (2011) and Zhang et al. (2014). The Xiliugou Formation is composed of fluvial-alluvial red sandstone with an approximate age of ~54-~33 Ma. The Yehucheng Formation consists of red lacustrine-aeolian mudstone and fluvial

sandstone with gypsum at the base, with an approximate age of ~33-23.6 Ma. The Xianshuihe Formation is characterized by red- and light-brown-coloured aeolian clay and fluvial sandstone that contains five local mammalian faunas (Qiu et al., 1997) and dates to 23.6-~9.1 Ma. The Linxia Formation consists of red aeolian clay interbedded with several fluvial sandstone layers in the lower part and the Wuquan Gravel in the upper part, with ages of ~8.3-~3.5 Ma. The Quaternary loess on the top of the section formed within the last 1.4 Myr and unconformably overlies the Linxia Formation.

3. Sampling and analytical methods

Pedogenic samples were collected at least 30-50 cm below the section surface, avoiding weathered surfaces and visible calcite or gypsum veins to minimize the possible effects of diffusion and diagenesis (Cerling and Quade, 1993; Hough et al., 2011). A total of 297 pedogenic carbonate samples, including carbonate cements and pseudomycelia, were sampled from the aeolian sediments wherever possible. The sampling intervals in the Nanshan and Jiuzhoutai sections ranged from 3 to 5 m and that in the Fenghuangshan section ranged from 10 to 100 m (Fig. 2).

Carbonate samples were converted to CO₂ using dehydrated phosphoric acid under a vacuum at 70 °C for 1 h. Oxygen and carbon isotope ratios were measured using a MAT-253 mass spectrometer at the Key Laboratory of Western China's Environmental Systems (Ministry of Education), Lanzhou University. The measurement precision is generally <0.1‰ and was checked using NBS-19 ($\delta^{18}\text{O}=-2.20\text{‰}$, $\delta^{13}\text{C}=+1.95\text{‰}$, PDB standard) and NBS-18 ($\delta^{18}\text{O}=-23.2\text{‰}$, $\delta^{13}\text{C}=-5.1\text{‰}$, PDB standard) measurements after every nine samples. The

measurement precisions for oxygen and carbon isotope ratios were 0.089‰ and 0.086‰, respectively. All isotope ratios are given in ‰ relative to PDB.

4. Results

The $\delta^{18}\text{O}$ values range from -14.29‰ to -4.80‰, with a mean of -9.25‰ (n=297, $1\sigma=1.42\%$; Table 1; Fig. 2). They exhibit an overall positive trend from the bottom to the top of the section. A more negative trend, averaging -12.46‰ (n=15, $1\sigma=1.25\%$), is evident in the lowest part of the Xiliugou Formation from ~46.5 Ma to ~33 Ma. The $\delta^{18}\text{O}$ values increase progressively in the Yehucheng Formation (mean -10.19‰, n=25, $1\sigma=0.95\%$) and in the Xianshuihe Formation (mean -8.81‰, n=35, $1\sigma=0.90\%$) from ~33 Ma to ~9.1 Ma. By contrast, a decreased mean value of -9.57‰ (n=145, $1\sigma=0.90\%$) is observed in the Linxia Formation from ~8.3 Ma to ~3.5 Ma. A sharp increase to a mean of -7.92‰ (n=77, $1\sigma=1.01\%$) occurs in the Quaternary loess, spanning the last 1.4 Myr.

The $\delta^{13}\text{C}$ values range from -9.37‰ to 0.48‰ with a mean of -4.70‰ (n=297, $1\sigma=1.95\%$; Table 1; Fig. 2) and exhibit a weak positive shift from the bottom to the top of the section, with a remarkable shift occurring at ~3.5 Ma. The $\delta^{13}\text{C}$ values are high in the lowermost strata (mean -5.02‰, n=5, $\sigma=0.72$) and decrease by 1-4‰ in the Xiliugou Formation (mean -8.07‰, n=6, $1\sigma=0.82\%$) from ~40 Ma to ~33 Ma. The values increase to a mean of -5.62‰ (n=209, $1\sigma=0.80\%$) in the Yehucheng, Xianshuihe and Linxia Formations until ~3.5 Ma. Subsequently, a remarkable positive shift (mean -1.93‰, n=77, $1\sigma=1.41\%$) occurs in the Quaternary loess.

(Table 1)

(Fig. 2)

5. Discussion

5.1. Evaluation of the diagenetic effects

The recrystallization process during diagenesis can significantly alter the isotopic signals of carbonates, causing the original carbonate $\delta^{18}\text{O}$ values to be more negative due to the interaction of calcite with meteoric water at high temperatures (e.g., Kent-Corson et al., 2009). Nevertheless, several lines of evidence suggest that diagenetic alteration is insignificant in the case of the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records of the Cenozoic sequence from the Lanzhou Basin: 1) Numerous isotopic studies of more than 5000 m of Cenozoic sediments buried at relatively shallow depths on the margin of the northern Tibetan Plateau revealed that the effect of diagenetic alteration on the carbonate $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ signals was minor (Graham et al., 2005; Kent-Corson et al., 2009; Zhuang et al., 2011). Compared to these records, the 3200-m-thick Cenozoic strata in the Lanzhou Basin indicate a shallower burial depth, suggesting that diagenesis has had a minor influence on the isotopic record. 2) Throughout the entire Cenozoic sequence, the carbonate $\delta^{18}\text{O}$ record of the Lanzhou Basin does not exhibit extremely low $\delta^{18}\text{O}$ values, suggesting that the isotopic signals have not been affected by diagenesis. 3) Diagenesis would likely homogenize the original carbonate isotopic signal (Quade et al., 2007), resulting in a strong positive covariation between the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values (Singh et al., 2012). However, the occurrence of high-frequency

changes and weak covariation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values from the Lanzhou Basin carbonates (Fig. 2 and 3) suggest that significant diagenetic alteration of the isotopic values has not occurred.

(Fig. 3)

5.2. Oxygen isotope record and its implication for aridity of the Lanzhou Basin

5.2.1. Evaluation of the oxygen isotope signal

The $\delta^{18}\text{O}$ value of pedogenic carbonate ($\delta^{18}\text{O}_{\text{pc}}$) is mainly determined by the $\delta^{18}\text{O}$ of the soil water ($\delta^{18}\text{O}_{\text{sw}}$), which in turn is strongly related to the $\delta^{18}\text{O}$ of local meteoric water ($\delta^{18}\text{O}_{\text{mw}}$) (e.g. Cerling, 1984; Quade et al., 1989b). Measurements of modern soils provide support for this relationship, indicating that $\delta^{18}\text{O}_{\text{pc}}$ values are positively correlated with $\delta^{18}\text{O}_{\text{mw}}$ values (Cerling, 1984; Cerling and Quade, 1993).

Further analysis revealed that the factors influencing $\delta^{18}\text{O}_{\text{sw}}$ values principally include temperature and evaporation of soil water. Cerling and Quade (1993) found that soil water temperature may affect the $\delta^{18}\text{O}_{\text{pc}}$ values in carbonate precipitated in equilibrium with soil water, but it makes only a small contribution to $\delta^{18}\text{O}_{\text{pc}}$ variation. The relationship between the isotopic fractionation and soil water temperature was defined as $1000 \ln \alpha_{\text{calcite-water}} = 18.03(10^3 T^{-1}) - 32.42$ by Kim and O'Neil (1997), where T , the response temperature, is specified in K, and $\alpha_{\text{calcite-water}}$ is the isotope equilibrium fractionation between calcite and soil water. According to the equation, a large change in soil temperature only produces a small change in $\delta^{18}\text{O}_{\text{pc}}$ (~ -0.22 to $-0.24\text{‰}/^\circ\text{C}$ in the range $0\text{--}30\text{ }^\circ\text{C}$) (Quade et al., 2007). Nevertheless, soil water evaporation, a fractionation process, can strongly enhance residual soil water $\delta^{18}\text{O}$,

causing an increase in $\delta^{18}\text{O}_{\text{pc}}$ values during the isotopic fractionation of carbonate minerals (Cerling and Quade, 1993; Quade and Cerling, 1995). This effect is significant in arid areas. For example, greater enrichment of $\delta^{18}\text{O}$ was found in bare soil compared to vegetated soil due to the greater evaporation in bare soil (Allison et al., 1984). Numerous studies have demonstrated that the effect of evaporative enrichment in $\delta^{18}\text{O}_{\text{pc}}$ is more significant than that of other factors (Quade et al., 1989b; Quade and Cerling, 1995; Stern et al., 1997). Therefore, evaporation may be one of the dominant factors that determines $\delta^{18}\text{O}_{\text{pc}}$ in arid areas.

Modern precipitation observations suggest that $\delta^{18}\text{O}_{\text{mw}}$ values are mainly influenced by local air temperature, precipitation and rainfall source. In mid-to-high latitude regions, $\delta^{18}\text{O}_{\text{mw}}$ composition is positively correlated with mean annual air temperature (Dansgaard, 1964; Rozanski et al., 1993; Quade and Cerling, 1995), and the slope of this empirical relationship is defined as $0.69\text{‰}/^{\circ}\text{C}$ (Dansgaard, 1964) or $0.70\text{‰}/^{\circ}\text{C}$ (Rozanski et al., 1993). However, in tropical latitudes, where a temperature effect is not obvious, $\delta^{18}\text{O}_{\text{mw}}$ values are negatively correlated with rainfall amount, with an average depletion rate of $-1.5 \pm 0.2\text{‰}$ per 100 mm. This negative correlation is known as the “amount effect” and has been confirmed by many researchers (e.g. Yurtsever and Gat, 1981; Rozanski et al., 1993). The effect is also considered significant in monsoon areas (Singh et al., 2012). Rainfall source changes can also affect $\delta^{18}\text{O}_{\text{mw}}$ values. In general, precipitation in inland and high latitude areas has lower $\delta^{18}\text{O}_{\text{mw}}$ values because condensation and rainout during water vapour transport result in preferential loss of ^{18}O . This is called the “continental effect” and the

“latitude effect” (Dansgaard, 1964).

(Fig. 4)

Observations of modern precipitation reveal a strong positive relationship between $\delta^{18}\text{O}_{\text{mw}}$ and temperature in and around the Tibetan Plateau (Tian et al., 2001; Yao et al., 2013), as well as a strong negative relationship between $\delta^{18}\text{O}_{\text{mw}}$ and precipitation amount in the monsoon region (Yao et al., 2013). Nevertheless, the International Atomic Energy Agency (IAEA) data for Lanzhou and Zhangye (the Hexi Corridor) demonstrate that the $\delta^{18}\text{O}_{\text{mw}}$ values are positively correlated with precipitation (IAEA/WMO, 2006; Fig. 4). This relationship is opposite that in monsoon regions and may possibly be attributed to intense evaporation and limited precipitation. Modern soil $\delta^{18}\text{O}$ values in north China are positively correlated with aridity (Fig. 5), indicating that in semi-arid and arid areas, strong evaporation is the main cause of regional enrichment in topsoil $\delta^{18}\text{O}$.

(Fig. 5)

From the abovementioned information, it can be concluded that the $\delta^{18}\text{O}_{\text{pc}}$ values of aeolian sediments in semi-arid and arid areas such as the Lanzhou Basin in northwestern China are mainly controlled by evaporation and precipitation.

The progressively increasing trend of $\delta^{18}\text{O}_{\text{pc}}$ values in the Lanzhou Cenozoic sequence may result from decreased precipitation and intense soil water evaporation, as well as from enhanced aridity in the Lanzhou Basin. This interpretation is based on the following. First, the positive $\delta^{18}\text{O}_{\text{pc}}$ trend is consistent with decreasing basin

temperature, as reflected by redness (a^*) values and by global cooling (Fig. 6). This finding contradicts modern precipitation observations (Dansgaard, 1964; Rozanski et al., 1993; Quade and Cerling, 1995; Yao et al., 2013) and indicates that temperature is not the main factor responsible for $\delta^{18}\text{O}$ enhancement. Second, increases in $\delta^{18}\text{O}_{\text{pc}}$ coincide with intervals of high magnetic susceptibility values and aeolian components in the strata (Fig. 6). These parameters reflect decreased precipitation and strengthened aridity. Third, topsoil studies demonstrate a positive correlation between $\delta^{18}\text{O}$ values and an aridity index (Fig. 5). Fourth, a similar positive $\delta^{18}\text{O}$ shift is common in records from northwestern China and is generally associated with increased regional aridity (e.g. Graham et al., 2005; Kent-Corson et al., 2009; Zhuang et al., 2011).

5.2.2. Aridity history of the Lanzhou Basin

The $\delta^{18}\text{O}_{\text{pc}}$ record from the Lanzhou Basin exhibits a dramatic positive shift beginning at approximately 33 Ma, with subsequent sharp increases at approximately 22 Ma and 3.5 Ma. However, it exhibits more negative values from ~9-~3.5 Ma (Fig. 6). These characteristics potentially provide valuable evidence for reconstructing the record of regional aridification.

From ~46.5 Ma to ~33 Ma, the lower $\delta^{18}\text{O}_{\text{pc}}$ values are coeval with the development of fluvial and alluvial strata with the lowest magnetic susceptibility values (Zhang et al., 2014; Fig. 6). Previous studies have confirmed that the aeolian content significantly determines the variation of the magnetic susceptibility of sediments in the Lanzhou Basin because different types of sediment have

characteristic ranges of susceptibility values, with higher values in the aeolian sediments than in the fluvio-lacustrine sediments (Sun et al., 2011; Zhang et al., 2014). In addition, higher/lower magnetic susceptibility values reflect a drier/more humid environment. This is consistent with the lower $\delta^{18}\text{O}_{\text{pc}}$ values from ~46.5–~33 Ma, suggesting the occurrence of significant precipitation and a relatively humid climate. Evidence for the widespread occurrence of humid environments, as revealed in records from adjacent arid areas (e.g. Licht et al., 2014; Wang et al., 2014) and on a global basis (Zachos et al., 2001; Cramer et al., 2009), underlines the significance of this climatic interval.

(Fig. 6)

At ~33 Ma, the $\delta^{18}\text{O}_{\text{pc}}$ values exhibit a sharp positive shift, which is consistent with the increase in the aeolian contents and specifically with an increase in the long-term suspension dust (LSD) components (Fig. 6), indicates the onset of aridity in the Lanzhou Basin. This aridification process has been widely recorded across the East Asian continent, e.g. by significant increases in xerophytic plants, including *Ephedra*, in the Xining (Wang et al., 1990), Qaidam (Wang et al., 1999) and Lanzhou Basins (Miao et al., 2013); by a dramatic increase in the aeolian contents of sediments from the southwest Tarim Basin (Wang et al., 2014); and by the abrupt replacement of large perissodactyl-dominant faunas by small rodent faunas in Mongolia (Meng and McKenna, 1998). This regional aridification process suggests that the Asian inland environment underwent a fundamental change at this time. This sharp Eocene-Oligocene climatic transition may be attributed to the westward retreat of the

Tethys Sea during the late Eocene, with the amount of moisture transported by westerly winds gradually being reduced (Bosboom et al., 2011; Sun and Jiang, 2013; Licht et al., 2014; Wang et al., 2014), and to the global climatic shift towards icehouse conditions (Zachos et al., 2001; Dupont-Nivet et al., 2007; Abels et al., 2011; Licht et al., 2014).

From ~22 Ma to ~9.1 Ma, the remarkable positive shift in $\delta^{18}\text{O}_{\text{pc}}$, together with the dramatic increase in magnetic susceptibility, reflects an intensification of aridification in the Lanzhou Basin. This event is coeval with the development of large-scale aeolian strata in the western CLP (Guo et al., 2002; Qiang et al., 2011; Zhang et al., 2014), with the increase in the dust flux in the North Pacific Ocean (Rea et al., 1985), and with the positive shift in carbonate $\delta^{18}\text{O}$ values from the northern margin of the Tibetan Plateau (Kent-Corson et al., 2009). These broadly consistent records suggest large-scale aridification occurred in the Asian interior since the early Miocene. Although the relatively low magnetic susceptibility values and aeolian contents of the sediments from ~17–~14 Ma may be a response to the relatively warm Middle Miocene Climate Optimum (MMCO), the relatively high $\delta^{18}\text{O}_{\text{pc}}$ values during this period suggest an overall arid climatic pattern. This early Miocene regional aridification process was likely causally related to the accelerated uplift of the Tibetan Plateau since the late Oligocene (e.g. Yin, 2006; DeCelles et al., 2007; Royden et al., 2008; Lu and Xiong, 2009; Molnar and Stock, 2009; Xu et al., 2013), which blocks water vapour pathways and results in decreased precipitation over the northeastern Tibetan Plateau. In addition, the shrinking of the Paratethys from the late Oligocene to

the Miocene (Meulenkamp and Sissingh, 2003; Popov et al., 2004; Fig. 6) likely contributed to this drying event (Guo et al., 2002).

From ~8.3 Ma to ~3.5 Ma, the remarkable negative shift in $\delta^{18}\text{O}_{\text{pc}}$ values, together with lower magnetic susceptibility and higher redness (a^*) values, indicates a relatively warm, humid environment in the Lanzhou Basin. The values of $\delta^{18}\text{O}_{\text{pc}}$, redness and magnetic susceptibility are similar to those of the late Oligocene, suggesting similar climatic conditions during both intervals. This environmental transition from arid to relatively wet conditions occurred against a background of Tibetan Plateau uplift (Li et al., 2014; Zhang et al., 2014) and global cooling (Zachos et al., 2001; Cramer et al., 2009). This was likely due to the low elevation (probably less than 1000 m) of the basin, when the sediments of the Linxia Formation accumulated (Sun et al., 2011b). A model for this interpretation is that the high elevation of the FHS section resulted from tectonic uplift that occurred at 8-9 Ma, causing a cessation of the Xianshuihe Formation accumulation in the northwestern part of the Lanzhou Basin; however, the NS section, to the southeast of the basin with a low elevation, began to accumulate Linxia Formation after ~8.3 Ma (Zhang et al., 2014). Nevertheless, the increase in $\delta^{18}\text{O}_{\text{pc}}$ values after 7.2 Ma, corresponding to a progressive increase in the aeolian contents and magnetic susceptibility values, suggests that an arid environment re-developed in the basin. The synchronous accumulation of red clay in the central CLP (Sun et al., 1998a, 1998b; Ding et al., 1999) and the increase of dust flux and grain size in North Pacific Ocean sediments (Rea et al., 1985) suggest a significant expansion of the arid area within the Asian

continent. In addition, the stepwise enhancement of $\delta^{18}\text{O}_{\text{pc}}$ values, together with the sharp increase in the magnetic susceptibility from 5.3 Ma, indicates a further intensified aridification process within the Lanzhou Basin. This process is widely observed on the northeastern margin of the Tibetan Plateau. For example, there were abrupt increases in xerophytic plants in the Qaidam (Wang et al., 1999), Tarim (Zhang and Sun, 2011) and Jiuxi Basins (Ma et al., 2005), as well as the terminal of lake environment of Lop Nor in the Tarim Basin (Liu et al., 2014). These two drying events are generally considered the consequence of significant Tibetan Plateau uplift during the late Miocene (Sun et al., 1998a, 1998b; Ding et al., 1999; Sun et al., 2008; Zhang and Sun, 2011; Liu et al., 2014). However, the enhanced aridity at 5.3 Ma may partly have been a result of the Mediterranean salinity crisis (Hsü et al., 1977; Krijgsman et al., 1999; Ryan, 2009), which significantly reduced the moisture carried by the westerly circulation, causing a precipitation decrease in the Asian interior.

At ~3.5 Ma, the $\delta^{18}\text{O}_{\text{pc}}$ values exhibit a dramatic positive shift, with the highest values in the entire sequence occurring within the Quaternary loess. This reflects further intensification of aridification in the Lanzhou Basin. This enhanced drying process is widely recorded in northwestern China, e.g., by the dramatic increases in xerophytic plants in the Qaidam (Wu et al., 2011), Tarim (Cao et al., 2001) and Jiuxi Basins (Ma et al., 2005) at ~3.5 Ma and by the formation of the Taklimakan Desert in the Tarim Basin at ~3.4 Ma (Sun et al., 2011a). Furthermore, the sediment accumulation rates at numerous sites on the CLP and the dust flux into the North Pacific Ocean both increased significantly at the same time (e.g. Rea et al., 1985,

1998; Sun et al., 1998a, 1998b). These features suggest enhanced aridification in inland Asia. Tibetan Plateau uplift is generally regarded as the major driving force for this palaeoenvironmental shift (An et al., 2001; Sun et al., 2011a). This phase of tectonic activity, namely the Qingzang Movement (Li, 1991), widely and profoundly affected the geomorphic and geological patterns of the Tibetan Plateau. For example, extensive molasse deposits began to develop in the Qilian, Kunlun and Tianshan Mountains (Chen et al., 2002; Wang et al., 2003; Sun J. et al., 2004; Huang et al., 2006), and the sediment accumulation rate, grain size and conglomerate contents of sediments increased dramatically in the intermontane basins (Jiuquan, Qaidam, Guide, Linxia and Tianshui Basins) (Li et al., 2014). In addition, the development of the loess-palaeosol sequence was a response to the development of high amplitude global glacial and interglacial climatic cycles during the Quaternary.

5.3. Carbon isotope record and its implication for palaeoecological change in the Lanzhou Basin

5.3.1. Factors influencing carbon isotopic composition

The $\delta^{13}\text{C}$ of pedogenic carbonate ($\delta^{13}\text{C}_{\text{pc}}$) is related to the $\delta^{13}\text{C}$ composition of soil CO_2 , which is mainly controlled by the relative biomass proportions of C_3 , C_4 and CAM plants where the soil respiration rates are high (Cerling, 1984, 1991; Cerling and Quade, 1993). These three plant groups have characteristic $\delta^{13}\text{C}$ values, ranging from -37‰ to -20‰ with a mean of -27‰ for C_3 plants, -14‰ to -10‰ with a mean of -12‰ for C_4 plants and intermediate values for CAM plants, which typically constitute a less significant proportion of the biomass (Rabenhorst et al., 1984; Kohn,

2010). Growing-season temperature, the partial pressure of CO₂ ($p\text{CO}_2$), light intensity, and soil water conditions are the main factors that determine the composition of the plant community via their effects on photosynthetic efficiency (Cerling and Quade, 1993; Ehleringer et al., 1997). C₄ plants have a greater photosynthetic efficiency than C₃ plants at low $p\text{CO}_2$ (roughly less than 500 ppmV), high growing-season temperature, high light intensity and enhanced summer precipitation conditions (Cerling and Quade, 1993; Ehleringer et al., 1997; Sage et al., 1999). Moreover, C₄ plants have higher water-use efficiencies than C₃ plants (Sage et al., 1999). This is one of the main reasons for the widespread distribution of C₄ plants in areas with seasonal and arid climates (e.g. An et al., 2005; Sun et al., 2013). In addition, C₃ plants are enriched in ¹³C under water-stressed conditions, resulting in C₃ plants that grow in arid environments having more positive $\delta^{13}\text{C}$ values compared to those that grow in less water-stressed conditions (Ehleringer, 1989; Kohn, 2010). Based on the specific $\delta^{13}\text{C}$ composition of the vegetation, $\delta^{13}\text{C}_{\text{pc}}$ represents different combinations of plant biomass. In general, $\delta^{13}\text{C}_{\text{pc}}$ values of $\sim +2\text{‰}$ represent a C₄-dominated community, whereas values of $\sim -12\text{‰}$ represent a C₃-dominated community, and values between -7‰ and $+2\text{‰}$ represent a mixed and/or water-stressed community (Ehleringer and Cerling, 2002).

5.3.2. Inferred palaeo-vegetation evolution

It is generally accepted that the expansion of C₄ plants occurred since the late Miocene (e.g. Quade and Cerling, 1995; Cerling et al., 1997; Latorre et al., 1997; Ding and Yang, 2000). This trend is also recorded in the Cenozoic sequence in the

Lanzhou Basin.

The $\delta^{13}\text{C}_{\text{pc}}$ record exhibits high values, averaging $-1.93 \pm 1.41\text{‰}$ in the Quaternary loess sediments (Fig. 6). These high $\delta^{13}\text{C}_{\text{pc}}$ values can be explained by three possible mechanisms: 1) the contribution of detrital carbonate; 2) the expansion of C_4 plants; and 3) the consequence of sparse local vegetation. We suggest that vegetation density may have played a significant role, and change in C_3/C_4 ratio made an important contribution.

Carbonate in aeolian sediments generally comprises detrital carbonate from palaeo-marine sediments in the source region, which have very high $\delta^{13}\text{C}$ values that could enhance the $\delta^{13}\text{C}$ values of the total carbonate (e.g. Frakes and Jianzhong, 1994; Li and Liu, 2003; Liu et al., 2005; Ning et al., 2006), especially in the arid northwestern CLP due to the weak soil leaching as result of low precipitation (Liu et al., 1985). However, several lines of evidence suggest that detrital carbonate has limited contribution to the $\delta^{13}\text{C}_{\text{pc}}$ of the Cenozoic sequence in the Lanzhou Basin. First, the content of detrital carbonate compared to the total carbonate content in aeolian sediments is quite limited, generally less than 10% (e.g. Wen et al., 1989; Frakes and Jianzhong, 1994; Li and Liu, 2003). This detrital carbonate content would not significantly affect the $\delta^{13}\text{C}$ values of the total carbonate. Second, the $\delta^{18}\text{O}$ record of the Quaternary loess from the Lanzhou Basin exhibits a similar pattern of variation as those from the central and southern CLP (e.g. Ding and Yang, 2000; An et al., 2005; Yang et al., 2012), suggesting that they were consistently influenced by post-depositional pedogenesis. Third, high $\delta^{13}\text{C}_{\text{pc}}$ values were observed in several

records from the western CLP (Dettman et al., 2003; Rao et al., 2006; Liu et al., 2011), which are believed to have been significantly affected by post-depositional pedogenesis driven by monsoonal precipitation. Compared with these records, a similar range of $\delta^{13}\text{C}_{\text{pc}}$ variation in the Lanzhou Basin suggests a consistent controlling mechanism and that detrital carbonate has a negligible influence on the isotopic record.

Due to the distinct $\delta^{13}\text{C}$ ranges of C_3 and C_4 plants, the $\delta^{13}\text{C}_{\text{pc}}$ variations in loess-palaeosol sequences have been explained as a function of the C_3/C_4 plant ratio of the local vegetation (e.g. Cerling, 1984; Quade et al., 1989b). Research regarding the main CLP has proposed that the increasing $\delta^{13}\text{C}_{\text{pc}}$ values reflected the abundance of C_4 plants within the local ecosystem (e.g. Ding and Yang, 2000; Jiang et al., 2002; An et al., 2005; Kaakinen et al., 2006; Sun et al., 2012). The $\delta^{13}\text{C}_{\text{pc}}$ variations in aeolian sediments from the western CLP were also interpreted as the result of variations in the ratios of C_3 and C_4 plants (e.g. Wang et al., 2005; Hough et al., 2011). The $\delta^{13}\text{C}_{\text{pc}}$ values in the late Cenozoic sequence in the Lanzhou Basin fall within the isotopic composition range associated with a mixed community consisting of C_3 and C_4 plants (Ehleringer and Cerling, 2002), and the $\delta^{13}\text{C}$ records of bulk organic matter from the Lanzhou Basin imply the presence of C_4 plants during the late Cenozoic (Yang et al., 2015). Thus, we believe that the change in C_3/C_4 plant ratio may have significantly contributed to the $\delta^{13}\text{C}_{\text{pc}}$ variations in the Lanzhou Basin, and that the increase in the abundance of C_4 plants in the local ecosystem may explain the remarkable positive trend in $\delta^{13}\text{C}$ record after ~3.5 Ma in the Quaternary loess in the

Lanzhou Basin.

However, it is difficult to believe that all of the $\delta^{13}\text{C}_{\text{pc}}$ variations were controlled by the C_3/C_4 plant ratio, and it seems likely that at least some of the variations reflect other factors. This can be illustrated by spatial and temporal changes in $\delta^{13}\text{C}_{\text{pc}}$ values. In the temporal domain, $\delta^{13}\text{C}_{\text{pc}}$ values in high-resolution records generally exhibit high frequency fluctuations with durations of several thousand years (e.g. An et al., 2005; Rao et al., 2006). The alteration of the C_3/C_4 ratio in such a short time would require unrealistically rapid alternation between completely different plant groups. Such a scenario is likely impossible from an ecological viewpoint because vegetation contraction and recovery are generally associated with gradual evolution or modification rather than complete turnover from one plant group to another. A similar situation occurs in the case of the spatial variability of $\delta^{13}\text{C}_{\text{pc}}$. For example, $\delta^{13}\text{C}_{\text{pc}}$ variations in the loess-palaeosol sequence during the last glacial cycle in the CLP exhibit significant variability between different sections that are not widely separated (e.g. Jiang et al., 2001). It is difficult to understand how one site could have had a vegetation type with a particular C_3/C_4 ratio while another site several kilometres away had a different vegetation type and a different C_3/C_4 ratio. Such variability in $\delta^{13}\text{C}_{\text{pc}}$ values may be the consequences of local differences in vegetation density related to local environmental factors such as topography, soil humidity and soil nutrient levels. Therefore, we suggest that local factors, particularly vegetation density, may have played an important role in determining some of the $\delta^{13}\text{C}_{\text{pc}}$ variations.

Evidence exists that supports the effects of vegetation density on $\delta^{13}\text{C}_{\text{pc}}$ variations in the aeolian sediment. First, vegetation density can affect the soil CO_2 concentration, with lower vegetation density yielding a lower soil CO_2 concentration (Amundson et al., 1988; Stevenson et al., 2005). A low soil CO_2 concentration increases the contribution of atmospheric CO_2 during pedogenic carbonate formation, increasing the $\delta^{13}\text{C}_{\text{pc}}$ values because atmospheric CO_2 is more enriched in ^{13}C than soil CO_2 (e.g., Cerling and Quade, 1993). Second, $\delta^{13}\text{C}_{\text{pc}}$ variations are opposite C_3/C_4 ratio variations and are consistent with the temporal trend in vegetation change. Studies regarding the $\delta^{13}\text{C}$ values of organic matter in loess-palaeosols demonstrated that the C_4 biomass increased consistently from glacials to interglacials (e.g. Liu et al., 2005; Yang et al., 2015). However, the $\delta^{13}\text{C}_{\text{pc}}$ variation exhibits an opposing trend and was more positive in the loess (glacials) than in the palaeosol (interglacials) layers (e.g. Rao et al., 2006; Liu et al., 2011). This cannot be explained by the C_3/C_4 ratio and is consistent with changes in vegetation density. Third, a similar situation was observed for the spatial $\delta^{13}\text{C}_{\text{pc}}$ decrease from the northwestern to southeastern CLP in both the loess and palaeosol layers (e.g., Rao et al., 2006). This trend is opposite that of the C_3/C_4 ratio variation and consistent with vegetation density changes on the CLP. The $\delta^{13}\text{C}_{\text{pc}}$ values in the Quaternary loess sequence from the Lanzhou Basin demonstrate this spatial attribute.

6. Conclusions

A near-continuous record of Cenozoic $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ variations in pedogenic carbonates in the Lanzhou Basin, Northwestern China reveals the history of Asian

inland aridification and the pattern of local palaeovegetation. Before ~33 Ma, the $\delta^{18}\text{O}$ values indicate a relatively humid climate. Subsequently, the $\delta^{18}\text{O}$ values became more positive with distinct positive shifts at ~33, ~22 and ~3.5 Ma, suggesting the onset of aridification at ~33 Ma that further intensified at ~22 Ma and ~3.5 Ma. These trends are generally synchronous with those of other proxy records from the region and reveal the process of progressive aridification across the Asian interior during the Cenozoic. The three step shifts are closely linked to Tibetan Plateau uplift, westward retreat of the Tethys Sea and ongoing global cooling. The $\delta^{13}\text{C}$ values exhibit a weak positive trend with a remarkable shift at ~3.5 Ma. This trend is likely related to a decrease in vegetation density in response to Cenozoic drying. The remarkable shift at ~3.5 Ma may be explained by the large-scale expansion of C_4 plants.

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Figure captions

Fig. 1. Distribution of aeolian sediments in northern China (modified from Zhang et al., 2014). The dashed black line indicates the limit of the Asian summer monsoon identified by Gao (1962).

Fig. 2. Stratigraphy, geomagnetic polarity, lithology, chronology (Qiu et al., 1997; Sun et al., 2011a; Zhang et al., 2014) and oxygen and carbon isotopic records from the studied sections in the Lanzhou Basin.

Fig. 3. $\delta^{18}\text{O}$ versus $\delta^{13}\text{C}$ of pedogenic carbonates from the studied sections in the Lanzhou Basin.

Fig. 4. Mean monthly temperature, precipitation and $\delta^{18}\text{O}$ values of precipitation from Lanzhou and Zhangye obtained from the Global Network of Isotopes in Precipitation database (IAEA/WMO, 2006).

Fig. 5. Results of oxygen isotope studies of modern soils in north China (Ding and Yang, 2000; Wang et al., 2005; Sheng et al., 2008), with base map adapted from the Annual Aridity of China (Atlas of Physical Geography in China, 2ed, p44). The black contours represent the drought index (evaporation/precipitation); red circles represent soil sampling locations and the red numbers are isotopic values.

Fig. 6. Time series of oxygen and carbon isotopic composition since ~44.6 Ma compared with other palaeoenvironmental proxy indicators from the sections in the Lanzhou Basin (Zhang et al., 2014) and the deep sea oxygen isotope record (Cramer et al., 2009). A- $\delta^{18}\text{O}_{\text{pc}}$, B- $\delta^{13}\text{C}_{\text{pc}}$, C-redness (a^*), D-magnetic

susceptibility, E-aeolian content, F-long-term suspension dust (LSD) content, and G-the deep-sea oxygen isotope record. The final three columns show major global cooling events, and episodes in the uplift of the Tibetan Plateau and in the evolution of the Tethys Sea.

Table captions

Table 1 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of pedogenic carbonates from the studied sections in the
Lanzhou Basin

Figure

Fig. 1

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Fig. 3

Fig. 4

Fig. 5

Fig. 6

Table 1

$\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of pedogenic carbonates from the studied sections in the Lanzhou Basin

Sample name	Depth (m)	Age (Ma)	$\delta^{18}\text{O}$ (PDB, ‰)	$\delta^{13}\text{C}$ (PDB, ‰)
Jiuzhoutai Section				
<i>Quaternary loess</i>				
JZT001	5.47	0.01	-6.11	0.02
JZT002	9.47	0.01	-8.25	-0.29
JZT003	13.48	0.02	-8.34	-0.45
JZT004	17.48	0.02	-8.39	-0.92
JZT005	21.32	0.03	-8.07	-1.31
JZT006	25.50	0.03	-6.37	-3.27
JZT007	27.95	0.03	-8.06	-0.94
JZT008	29.63	0.04	-8.29	-0.65
JZT009	30.68	0.04	-8.78	-0.96
JZT010	30.73	0.04	-8.58	-1.14
JZT011	42.13	0.06	-4.80	-0.20
JZT012	49.46	0.07	-6.46	0.48
JZT013	57.63	0.13	-9.92	-6.41
JZT014	66.79	0.19	-8.04	-1.52
JZT015	71.56	0.22	-7.65	-2.46
JZT016	75.77	0.24	-8.48	-2.10
JZT017	80.44	0.27	-9.13	-0.87
JZT018	84.50	0.29	-8.58	-0.96
JZT019	90.25	0.32	-8.62	-4.92
JZT020	95.17	0.35	-5.94	-2.15
JZT021	96.23	0.35	-7.52	-5.59
JZT022	99.40	0.37	-10.54	-2.68
JZT023	102.71	0.39	-8.80	-0.85
JZT024	105.71	0.40	-8.52	-1.05
JZT025	109.09	0.41	-8.74	-1.94
JZT026	114.23	0.44	-5.53	-5.43
JZT027	117.97	0.45	-8.69	-0.63
JZT028	122.30	0.47	-8.97	-0.87
JZT029	127.07	0.49	-8.99	-1.08
JZT030	131.01	0.51	-8.63	-2.15
JZT031	134.55	0.53	-7.20	-0.52
JZT032	137.88	0.54	-8.45	-1.66
JZT033	141.02	0.56	-6.75	-3.35
JZT034	144.42	0.57	-7.63	-3.27
JZT035	148.56	0.59	-9.25	-5.02
JZT036	152.56	0.61	-6.77	-3.40
JZT037	156.00	0.62	-7.63	-1.64

JZT038	159.77	0.63	-5.69	-4.80
JZT039	163.59	0.65	-6.76	-2.69
JZT040	166.98	0.67	-8.49	-1.48
JZT041	170.60	0.68	-8.11	-0.60
JZT042	173.86	0.69	-7.72	-1.37
JZT043	177.22	0.71	-8.65	2.26
JZT044	180.81	0.72	-9.12	-1.57
JZT045	184.28	0.74	-9.04	-1.64
JZT046	187.55	0.75	-7.07	-4.75
JZT047	190.63	0.76	-8.07	-1.35
JZT048	194.71	0.78	-8.94	-0.68
JZT049	198.70	0.80	-8.53	-1.78
JZT050	202.78	0.82	-8.10	-2.02
JZT051	206.98	0.85	-8.04	-1.90
JZT052	210.57	0.86	-8.21	-3.59
JZT053	214.97	0.89	-7.75	-2.54
JZT054	219.08	0.91	-8.45	-0.52
JZT055	222.52	0.93	-7.71	-1.48
JZT056	226.82	0.95	-6.75	-3.31
JZT057	230.55	0.97	-7.02	-2.70
JZT058	234.57	0.99	-7.41	-3.95
JZT059	238.58	1.01	-8.39	-1.23
JZT060	242.60	1.03	-8.11	-1.03
JZT061	246.61	1.05	-7.93	-1.81
JZT062	251.20	1.08	-8.24	-1.31
JZT063	254.64	1.10	-8.17	-0.78
JZT064	258.66	1.12	-8.17	-1.63
JZT065	262.67	1.14	-8.34	-1.39
JZT066	266.69	1.16	-8.20	-1.09
JZT067	270.70	1.18	-8.38	-1.20
JZT068	274.72	1.20	-7.48	-3.79
JZT069	278.73	1.22	-8.27	-1.07
JZT070	282.75	1.25	-6.35	-1.35
JZT071	286.76	1.27	-5.88	-2.35
JZT072	290.78	1.29	-7.87	-2.02
JZT073	294.79	1.31	-7.91	-1.35
JZT074	298.81	1.33	-7.53	-2.15
JZT075	302.82	1.35	-7.82	-1.75
JZT076	306.84	1.37	-8.54	-0.93
JZT077	310.28	1.39	-7.45	-0.94
Nanshan Section				
<i>Linxia Formation</i>				
NS001	320.35	3.49	-9.91	-5.84
NS002	321.32	3.50	-9.41	-5.72

NS003	321.91	3.51	-9.67	-5.97
NS004	324.16	3.54	-9.16	-5.92
NS005	324.98	3.55	-9.37	-5.95
NS006	327.59	3.59	-9.37	-6.83
NS007	330.04	3.62	-7.80	-5.79
NS008	332.49	3.65	-8.85	-6.31
NS009	335.27	3.69	-8.98	-6.44
NS010	337.39	3.72	-9.05	-7.46
NS011	339.89	3.75	-9.37	-6.92
NS012	342.32	3.79	-8.45	-5.40
NS013	345.05	3.83	-9.97	-7.97
NS014	347.45	3.86	-8.60	-7.48
NS015	350.14	3.90	-9.37	-6.92
NS016	352.38	3.93	-9.19	-7.20
NS017	355.27	3.97	-8.90	-5.87
NS018	357.44	4.00	-10.57	-7.65
NS019	360.03	4.03	-9.89	-7.75
NS020	362.42	4.06	-9.33	-7.83
NS021	365.10	4.10	-8.67	-6.77
NS022	367.45	4.13	-8.63	-6.14
NS023	370.13	4.17	-9.17	-6.41
NS024	372.47	4.20	-8.53	-6.36
NS025	374.98	4.22	-9.06	-7.07
NS026	377.50	4.25	-7.66	-5.41
NS027	380.01	4.28	-7.49	-5.55
NS028	382.52	4.35	-7.86	-4.74
NS029	385.03	4.44	-8.51	-6.31
NS030	387.55	4.49	-9.31	-6.03
NS031	389.82	4.52	-8.98	-5.79
NS032	392.45	4.55	-8.59	-4.76
NS033	395.12	4.58	-9.29	-5.58
NS034	397.56	4.60	-8.71	-4.80
NS035	399.98	4.63	-9.52	-5.93
NS036	402.41	4.66	-9.15	-5.39
NS037	405.02	4.69	-8.44	-5.35
NS038	407.41	4.71	-8.84	-5.50
NS039	410.15	4.74	-8.75	-5.67
NS040	412.48	4.77	-8.67	-5.33
NS041	415.05	4.80	-8.13	-4.91
NS042	417.38	4.81	-8.90	-5.66
NS043	419.95	4.83	-9.61	-6.32
NS044	422.57	4.85	-9.18	-5.48
NS045	425.19	4.86	-8.73	-5.54
NS046	427.45	4.87	-8.84	-6.39

NS047	430.13	4.89	-8.48	-5.32
NS048	432.44	4.91	-8.90	-5.36
NS049	435.02	4.93	-9.44	-5.68
NS050	437.59	4.95	-8.41	-4.74
NS051	440.15	4.97	-8.06	-4.66
NS052	442.49	5.00	-8.72	-5.63
NS053	445.05	5.04	-8.18	-4.94
NS054	447.57	5.08	-7.74	-4.51
NS055	449.70	5.11	-9.15	-4.56
NS056	452.54	5.15	-10.89	-5.39
NS057	454.96	5.18	-10.62	-5.14
NS058	457.38	5.22	-10.80	-6.40
NS059	459.98	5.25	-10.65	-5.32
NS060	462.40	5.28	-10.26	-4.81
NS061	465.01	5.31	-9.61	-5.05
NS062	469.93	5.36	-9.54	-4.82
NS063	472.36	5.39	-10.16	-5.59
NS064	475.01	5.42	-9.70	-4.87
NS065	477.44	5.45	-9.22	-4.83
NS066	479.90	5.48	-9.96	-5.27
NS067	482.39	5.50	-9.77	-5.23
NS068	484.89	5.53	-10.65	-6.12
NS069	487.38	5.56	-10.44	-5.76
NS070	490.11	5.59	-10.55	-5.49
NS071	492.36	5.62	-10.58	-5.49
NS072	494.97	5.65	-10.88	-5.96
NS073	496.83	5.67	-10.21	-5.56
NS074	500.10	5.70	-10.72	-5.87
NS075	502.43	5.73	-10.61	-5.82
NS076	504.82	5.76	-10.67	-5.77
NS077	507.12	5.78	-10.24	-5.04
NS078	509.87	5.82	-10.22	-4.79
NS079	512.50	5.85	-10.05	-4.89
NS080	515.04	5.87	-10.56	-4.84
NS081	517.71	5.91	-10.48	-3.52
NS082	520.43	5.94	-10.72	-4.37
NS083	522.24	5.96	-11.10	-4.02
NS084	525.36	6.00	-10.29	-4.18
NS085	527.59	6.03	-10.05	-3.91
NS086	529.81	6.05	-10.63	-4.92
NS087	532.77	6.09	-10.60	-5.26
NS088	534.99	6.12	-10.87	-5.14
NS089	537.21	6.14	-10.73	-5.31
NS090	540.18	6.16	-10.13	-5.40

NS091	542.40	6.18	-10.36	-4.52
NS092	545.36	6.20	-9.79	-5.18
NS093	547.48	6.21	-10.66	-5.71
NS094	550.04	6.23	-10.49	-6.20
NS095	552.61	6.25	-10.12	-5.90
NS096	555.17	6.27	-11.13	-6.26
NS097	557.73	6.36	-9.67	-5.98
NS098	559.65	6.43	-10.79	-6.68
NS099	562.88	6.54	-11.59	-6.96
NS100	565.29	6.62	-10.31	-5.99
NS101	567.71	6.70	-10.80	-6.11
NS102	570.13	6.78	-10.70	-6.01
NS103	571.74	6.84	-10.45	-5.70
NS104	574.61	6.94	-9.54	-5.68
NS105	577.48	6.96	-9.04	-5.52
NS106	580.35	6.99	-8.34	-5.11
NS107	583.21	7.02	-9.84	-5.78
NS108	585.13	7.03	-9.05	-5.60
NS109	587.04	7.05	-9.21	-4.95
NS110	591.22	7.09	-9.65	-6.22
NS111	592.61	7.10	-9.85	-5.46
NS112	595.40	7.12	-10.00	-4.88
NS113	598.19	7.14	-9.94	-5.84
NS114	599.58	7.15	-10.08	-6.12
NS115	602.37	7.16	-10.03	-5.07
NS116	603.76	7.17	-10.52	-5.22
NS117	607.42	7.25	-11.18	-4.98
NS118	610.16	7.31	-10.98	-4.58
NS119	612.91	7.36	-10.21	-5.52
NS120	614.58	7.37	-10.54	-5.12
NS121	617.02	7.40	-9.90	-5.29
NS122	618.39	7.42	-10.00	-5.73
NS123	622.53	7.46	-8.95	-5.86
NS124	625.30	7.49	-11.29	-4.48
NS125	627.38	7.52	-9.09	-6.01
NS126	630.16	7.55	-9.27	-5.38
NS127	632.32	7.57	-8.92	-5.66
NS128	635.43	7.60	-10.07	-5.82
NS129	637.76	7.63	-9.82	-5.89
NS130	640.13	7.66	-9.35	-6.03
NS131	642.52	7.68	-9.14	-6.07
NS132	645.71	7.72	-8.76	-6.73
NS133	648.10	7.74	-9.20	-6.13
NS134	650.49	7.77	-9.10	-6.47

NS135	653.69	7.80	-8.03	-5.38
NS136	657.72	7.85	-9.15	-5.58
NS137	660.14	7.88	-9.20	-5.86
NS138	662.56	7.90	-8.69	-6.17
NS139	664.97	7.93	-9.11	-7.01
NS140	667.39	7.96	-9.42	-6.41
NS141	669.81	7.98	-9.76	-5.14
NS142	672.14	8.01	-8.79	-7.07
NS143	675.18	8.04	-8.18	-5.59
NS144	680.50	8.10	-7.48	-4.24
NS145	682.78	8.12	-8.84	-4.75

Fenghuangshan Section

Xianshuihe Formation

FHS001	705.10	9.18	-6.69	-4.87
FHS002	754.19	9.82	-8.21	-5.25
FHS003	765.91	10.11	-8.35	-5.96
FHS004	774.73	10.27	-8.04	-4.84
FHS005	795.82	10.79	-7.78	-4.54
FHS006	813.44	11.27	-7.82	-4.56
FHS007	827.36	11.75	-8.72	-5.41
FHS008	828.46	11.80	-8.93	-5.08
FHS009	842.77	12.12	-9.52	-6.29
FHS010	852.60	12.38	-7.88	-4.90
FHS011	867.77	12.88	-8.37	-5.45
FHS012	914.60	13.73	-9.43	-4.80
FHS013	928.52	13.99	-9.86	-6.70
FHS014	976.18	14.71	-9.57	-6.86
FHS015	998.21	15.04	-9.05	-4.88
FHS016	1028.93	15.44	-10.03	-6.47
FHS017	1053.96	15.78	-8.06	-4.38
FHS018	1085.53	16.20	-10.62	-5.38
FHS019	1092.34	16.29	-8.81	-5.49
FHS020	1151.52	16.91	-7.97	-4.33
FHS021	1181.76	17.23	-10.35	-6.00
FHS022	1216.66	17.85	-9.33	-5.65
FHS023	1227.99	18.03	-9.23	-5.26
FHS024	1233.61	18.12	-10.09	-6.01
FHS025	1249.19	18.36	-9.41	-5.92
FHS026	1293.92	19.13	-9.90	-5.65
FHS027	1308.33	19.35	-8.30	-4.76
FHS028	1345.76	19.93	-9.09	-5.51
FHS029	1390.95	20.71	-8.78	-5.53
FHS030	1396.30	20.76	-9.35	-6.22
FHS031	1408.90	20.94	-9.10	-5.34

FHS032	1421.43	21.38	-8.41	-5.02
FHS033	1451.18	21.87	-7.59	-4.90
FHS034	1458.48	21.93	-7.89	-5.06
FHS035	1481.63	22.10	-7.92	-4.53
FHS036	1548.24	22.61	-11.94	-6.66
FHS037	1597.24	22.98	-9.53	-5.21
FHS038	1613.64	23.10	-9.17	-4.79
FHS039	1632.19	23.24	-10.64	-6.95
FHS040	1641.47	23.31	-8.80	-6.51
FHS041	1651.90	23.42	-10.15	-7.41

Yehucheng Formation

FHS042	1667.48	23.62	-9.14	-6.43
FHS043	1667.48	23.63	-9.16	-5.83
FHS044	1724.86	24.37	-9.86	-5.05
FHS045	1729.52	24.44	-11.10	-6.18
FHS046	1757.41	24.84	-11.44	-5.11
FHS047	1775.78	25.18	-8.54	-3.73
FHS048	1777.59	25.20	-10.33	-4.60
FHS049	1785.73	25.29	-10.91	-6.93
FHS050	1793.83	25.37	-10.28	-7.01
FHS051	1910.37	26.81	-8.59	-4.87
FHS052	1937.53	27.31	-10.36	-4.84
FHS053	1941.50	27.61	-10.19	-5.37
FHS054	1945.47	27.78	-10.65	-5.23
FHS055	1960.19	28.74	-11.56	-5.08
FHS056	1970.83	29.44	-10.47	-5.37
FHS057	1977.92	29.90	-9.93	-5.72
FHS058	1992.10	30.74	-9.98	-6.22
FHS059	1997.41	31.13	-11.72	-6.33
FHS060	2006.27	31.61	-10.41	-4.59

Xiliugou Formation

FHS061	2016.91	32.26	-14.14	-5.10
FHS062	2024.00	32.75	-14.28	-5.40
FHS063	2030.40	33.22	-13.21	-5.84
FHS064	2059.43	33.93	-12.04	-5.84
FHS065	2092.13	34.49	-11.70	-7.49
FHS066	2164.24	35.79	-12.97	-8.42
FHS067	2221.34	36.79	-11.51	-7.66
FHS068	2261.18	37.53	-11.22	-8.37
FHS069	2298.65	38.25	-11.51	-9.37
FHS070	2374.96	39.53	-13.08	-7.10
FHS071	2515.25	42.06	-11.22	-4.65
FHS072	2606.47	43.68	-10.14	-4.67
FHS073	2627.44	44.02	-13.05	-4.98

FHS074	2688.29	45.13	-12.50	-4.51
FHS075	2763.36	46.44	-14.29	-6.27

Site location of the Lanzhou Basin

$\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records of pedogenic carbonate in the Lanzhou Basin