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Uplift-driven climate change at 12 Ma: a long $\delta^{18}O$ record from the NE margin of the Tibetan plateau

David L. Dettman [a,*], Xiaomin Fang [b,c], Carmala N. Garzione [d], Jijun Li [b]

[a] Department of Geosciences, University of Arizona, Tucson, AZ 85721, USA
[b] State Key Laboratory of Loess and Quaternary Geology, Institute of Earth Environment, CAS, Xi'an, Shanxi 710054, PR China
[c] MOE National Laboratory of Western China's Environmental Systems and College of Resources and Environment, Lanzhou University, Lanzhou, Gansu 730000, PR China
[d] Department of Earth and Environmental Sciences, University of Rochester, Rochester, NY 14627, USA

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Abstract

Carbonates from fluvial and lacustrine sediments were sampled from multiple measured sections in the Linxia basin of western China. Based on textural and mineralogical evidence, lacustrine carbonates are primary precipitates from lake water. A 29 million year record of the oxygen isotope composition of meteoric water is inferred from the $\delta^{18}O$ values of these carbonates. This inference is based on the most negative $\delta^{18}O$ values in the lake carbonates, which represent lake waters that have experienced the least evaporative enrichment. Carbonate $\delta^{18}O$ values, a proxy for rainfall $\delta^{18}O$, are $\sim -10.5‰$ throughout the interval of 29–12 Ma. At 12 Ma there is a shift to $-9‰$, a value that remains into the Pliocene. This implies a major reorganization of atmospheric circulation patterns and a shift to more arid conditions at the NE margin of the Tibetan plateau with the post-12 Ma system similar to that of today. The 12 Ma event may represent the time at which the Tibetan plateau achieves sufficient elevation to block the penetration of moisture from the Indian Ocean or south Pacific into western China. The period of greatest aridity is from 9.6 to 8.2 Ma, a time interval which agrees well with other climate records.

Keywords: Tibetan plateau; uplift; oxygen isotopes; lacustrine; Miocene

1. Introduction

The development of high elevation in the Tibetan plateau has been suggested as a major forcing mechanism of climate change in South Asia [1,2]. A wide range of indicators from the Himalayan foreland, Arabian Sea, and Indian Ocean suggest links between high topography and changes in climate, South Asian paleoecology, and ocean chemistry (e.g. [3–9]). The timing of uplift remains a topic of major disagreement and different lines of evidence have suggested different uplift histories. Recent sedimentological and stable isotopic studies have provided constraints on paleoelevation and paleoclimate of extensional basins in the southern part of the Tibetan plateau, suggesting that this region attained a high elevation by ~ 11

* Corresponding author.
E-mail address: dettman@geo.arizona.edu (D.L. Dettman).

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million years ago (Ma) [10,11]. Climate records from the Siwalik sediments of Pakistan and Nepal show evidence of increased aridity at ~9 Ma based on both oxygen and carbon stable isotope records [3,12], and this has often been used as evidence for the onset of monsoonal circulation and, by inference, the development of a high plateau. However, sub-annual stable isotope data from the Nepal Siwaliks suggest that a strong wet–dry seasonality, consistent with a strong monsoon climate, was present as early as 10.7 Ma [7]. On the northern margin of the Tibetan plateau, extremely arid conditions in Central Asia deserts are believed to have developed as a result of uplift of the Tibetan plateau [13,14], perhaps as early as 22 Ma [15]. The record of eolian dust flux from North Pacific drill cores has been suggested as a proxy for climate change in Central Asia, with an increase in deep-sea sedimentation rates and grain size at 3.6 Ma interpreted as indicating the onset of extremely arid conditions in Central Asia associated with uplift of the Tibetan plateau [16]. With the exception of paleoelevation estimates from extensional basins in the plateau [10], the wide range of paleoclimate indicators used to infer uplift history are derived from regions distal to the Tibetan plateau. In this paper, we present a long-term O and C isotope record from Linxia basin on the northeastern edge of the Tibetan plateau in an effort to understand climate change in this proximal region and its relationship to the growth history of the Tibetan plateau (Fig. 1).

2. Geologic setting

Linxia basin is located just north of the topographic front on the northeastern edge of the Tibetan plateau, southwest of the city of Lanzhou (Fig. 1). Magnetostratigraphy of Linxia basin deposits demonstrates that it is one of the longest unbroken terrestrial records extant; deposition was continuous throughout the last 29 Myr, with the exception of an unconformity in the proximal part of the basin between 4.5 and 3.6 Ma [17,18]. Subsidence rates, facies distributions, and patterns of stratigraphic thickening suggest that subsidence was flexurally driven, associated with shortening in the northeastern margin of the Tibetan plateau [18].

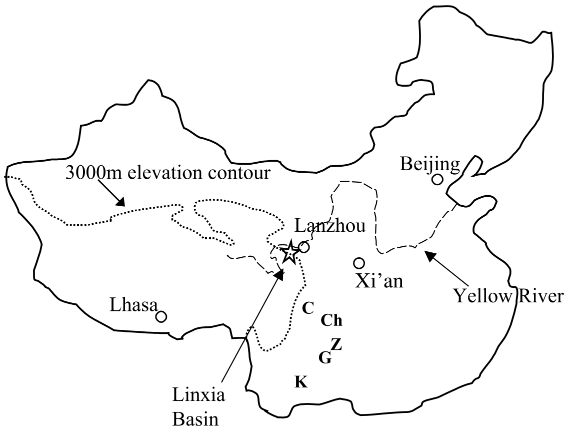

Fig. 1. Map of China, showing the location of Linxia basin relative to the Tibetan plateau. Letters mark locations of cities mentioned in text: C = Changqing; K = Kunming; Z = Zunyi, G = Guiyang; Ch = Chengdu.

The stable isotope record is based on carbonates in lacustrine and fluvial sediments from three stratigraphic sections in the Linxia basin (Fig. 2). In general, stratigraphic thickness ranges from over 1600 m proximal to the Tibetan plateau and pinches out on the flank of the Maxian Shan to the northeast. See Fang et al. [18] for a detailed description of the basin fill and its chronology. The sampled sections are located in the central part of the basin (Maogou section, 442 m thick) and proximal to the plateau margin (Wangjaishan and Dongshanding sections, > 1100 m thick). The paleomagnetic record used for the basis of our age control is from the Maogou section. This section records the beginning of deposition in the basin at 29 Ma to 4.3 Ma. The upper 500 m of the Wangjiashan section was also dated using paleomagnetic data. Here deposition continues from 11 Ma to the hiatus (4.5–3.6 Ma) and after the hiatus until 1.8 Ma. The Dongshanding section spans the last 2.4 Ma in 151 m, bringing the record into the Holocene.

The sediments filling the basin are made up of six fining upward cycles, which are identified as formations in the basin's stratigraphy [18]. Cycles start with sandstones or pebble conglomerates and grade into mudstone. In the central part of

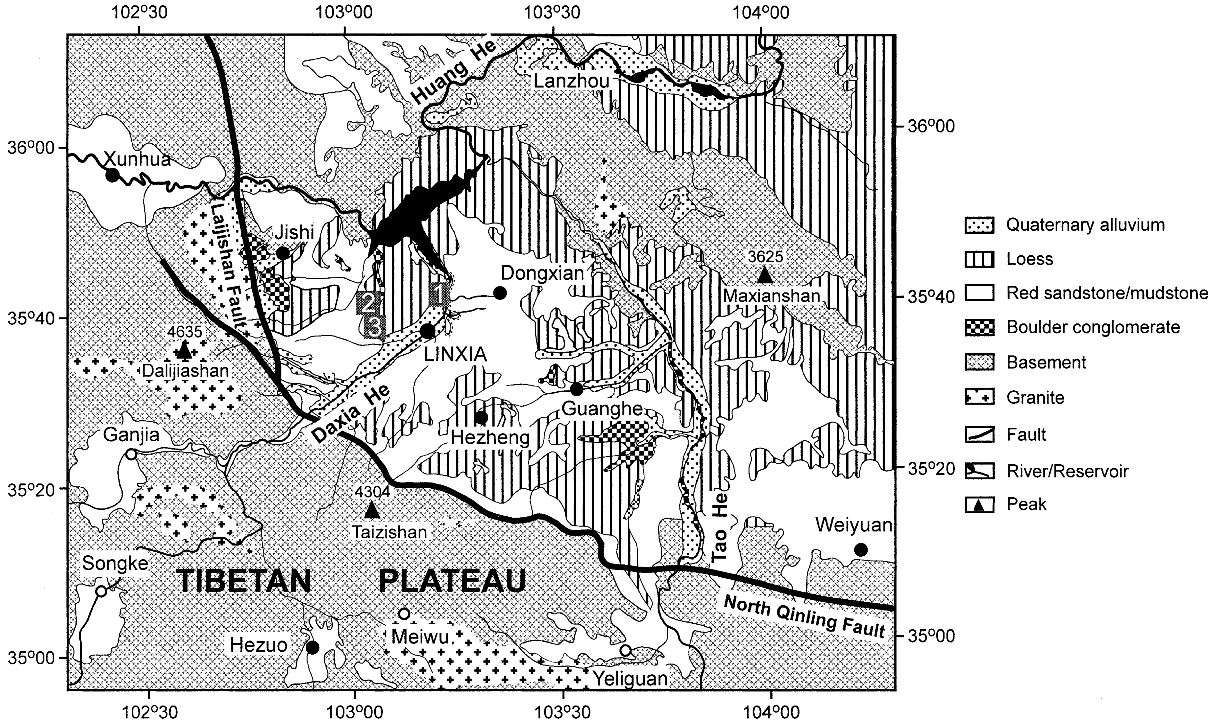

Fig. 2. Linxia basin surface geology and localities of measured sections: (1) Maogou section, (2) Wangjiashan section, (3) Dongshanding section.

the basin (Maogou section) fluvial sedimentation at the base of the section gives way to deltaic and lacustrine deposition, which persists from ~20 to ~7 Ma. In the Wangjiashan section, proximal to the plateau, sedimentary rocks are deformed and lacustrine deposition is much more frequently interrupted by fluvial deposits. After a depositional hiatus the uncomformable Jishi Formation, a boulder conglomerate, was deposited. Following that grayish/greenish mudstone, interpreted as subaqueous loess, to yellowish eolian loess of the Dongshanding Formation cap the Wangjiashan section. Younger loess deposits form a set of terraces along the Yellow River and Daxia River [17,18].

Fluvial and loess deposits are host to carbonates that are derived from fluvial cementation (e.g. [19]), soil processes, or from shallow groundwater cementation of sediments [20]. Soil carbonates are not common in the basin sediments and are primarily limited to the post-2.6 Ma loess deposits. Detrital carbonates are rare in sandstone thin sections (occasionally up to 2% of clast count), and are a very minor component in comparison to cementing carbonate.

The lacustrine deposits are dominantly red, non-calcareous mudstone. Carbonates occur in discrete clay-rich beige to greenish white to bright white beds on the order of 2–20 cm thick (Fig. 3). These calcareous mudstone layers are laterally extensive and follow bedding planes, and occasionally contain moldic fossils of aquatic organisms (very poorly preserved ostracodes and gastropods), which suggest that the carbonate precipitated from lake water. The mineralogy of these carbonate layers ranges from dolomite to Mg-rich calcite to low-Mg calcite.

3. Analytical methods

$\delta^{18}O$ and $\delta^{13}C$ of carbonates were measured using an automated carbonate preparation device (KIEL-III) coupled to a gas-ratio mass spectrom-

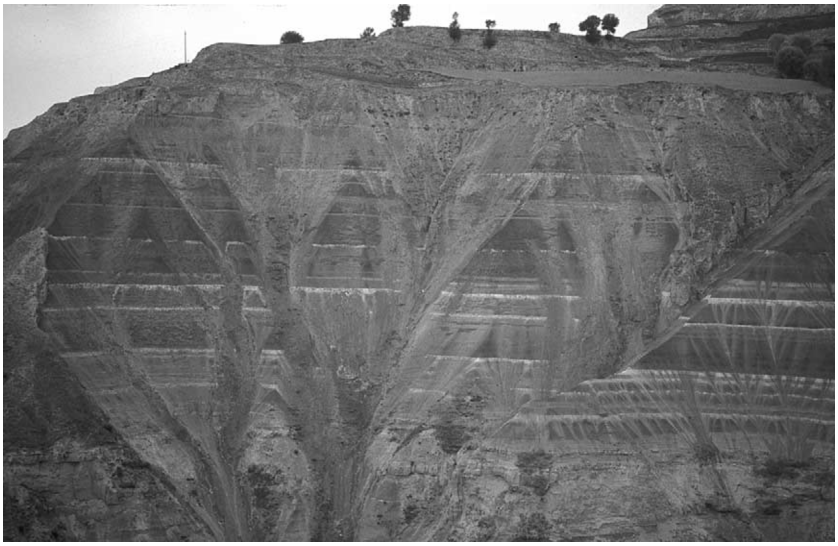

Fig. 3. Lacustrine deposits between 13 and 7 Ma. Light beds are calcareous, whereas darker intervals are non-calcareaous or mildly calcareous.

eter (Finnigan MAT 252). Powdered samples between 20 and 150 μg were reacted with dehydrated phosphoric acid under vacuum at 70°C. The isotope ratio measurement is calibrated based on repeated measurements of NBS-19 and NBS-18 and precision is ±0.1‰ for $\delta^{18}O$ and ±0.06‰ for $\delta^{13}C$ (1σ). Samples were heated under vacuum to 200°C prior to measurement.

Water $\delta^{18}O$ was measured using an automated CO_2-H_2O equilibration unit coupled to a Delta-S (Finnigan MAT) mass spectrometer. Standardization is based on internal standards referenced to VSMOW and VSLAP. Precision is better than ±0.08‰ for $\delta^{18}O$ (1σ).

Carbonate mineralogy was determined by X-ray diffraction based on the position of the d(104) peak [21]. Modern carbonates were used as representative end members for low-magnesium calcite (LMC; fresh-water mollusc shell), high-magnesium (HMC; echinoderm shell), and dolomite. LMC is defined as less than 4 mol%. HMC is defined as greater than 4 mol%.

4. Results

The record presented here (Fig. 4) is a composite based on samples from the Maogou section (29–4.6 Ma), the Wangjiashan section (3.6–1.8 Ma) and the Dongshanding section (1.8 Ma to present). Fluvial carbonates from 29 to 20 Ma have a mean $\delta^{18}O$ value of ∼−10.5‰ (VPDB). At 20 Ma lacustrine deposition begins and $\delta^{18}O$ values range from −11‰ to −5‰. After 13 Ma the $\delta^{18}O$ range of lake carbonates shifts to ∼−9‰ to a maximum value of −2.8‰ (VPDB). Although some variation exists, the $\delta^{18}O$ of fluvial carbonate immediately after 7 Ma maintains the $\delta^{18}O$ value of the most negative lake carbonates. After ∼3.6 Ma the oxygen isotope variability in loess deposits is greater than other intervals of fluvial deposition. Carbon isotope ratios tend to be more variable than oxygen isotope ratios during fluvial intervals and less variable during the lacustrine intervals. The $\delta^{13}C$ values of carbonates often move in parallel to changes in $\delta^{18}O$ during the lacustrine intervals.

5. Controls on carbonate $\delta^{18}O$

5.1. Diagenesis

A number of features suggest that these carbon-

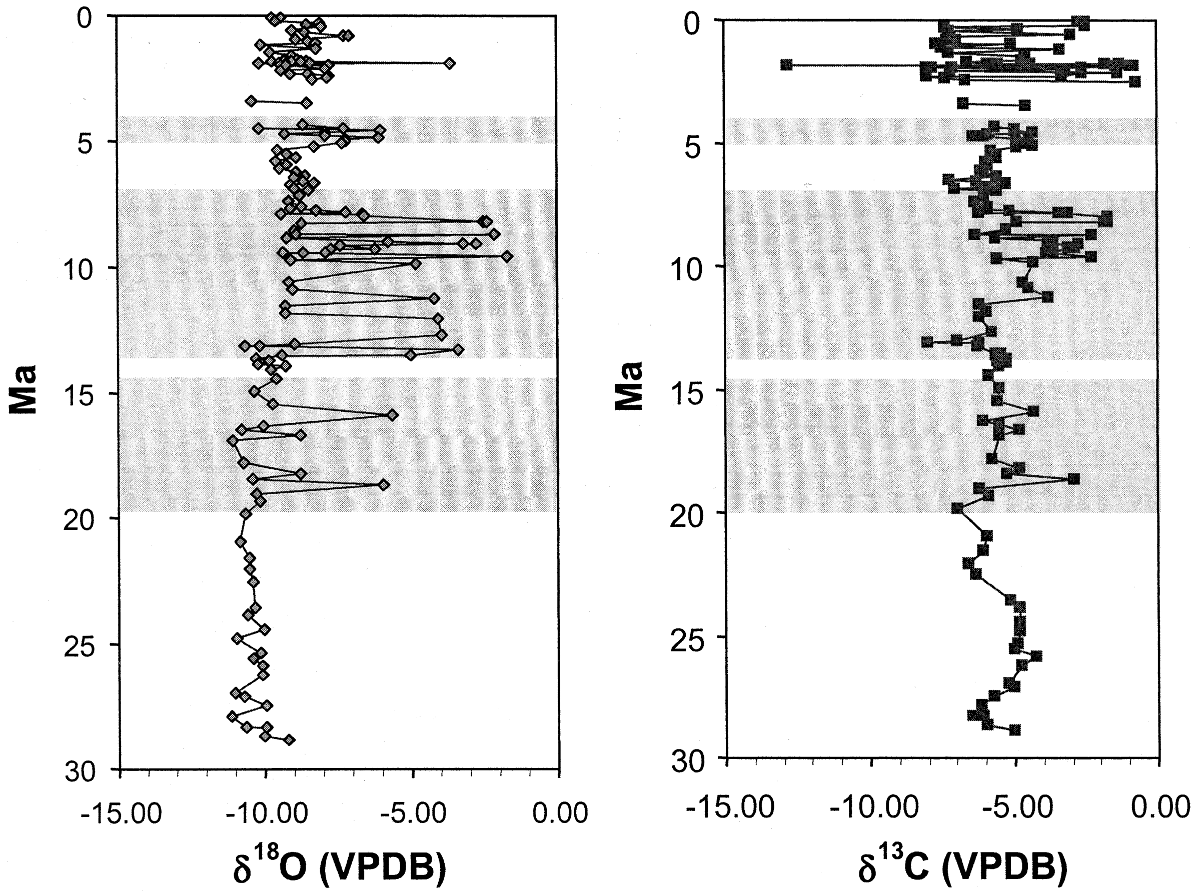

Fig. 4. Oxygen and carbon isotope composition of Linxia basin carbonates. Shaded portions are lake deposits. Chronology is based on paleomagnetic dating [17,18].

ates have not experienced significant diagenesis. First, the textural evidence noted above argues for a primary lacustrine precipitate. Secondly, there is no evidence of later remobilized carbonate in sediments immediately adjacent to these beds, as sediments above and below are free of carbonate. Finally, based on X-ray diffraction data, there is a strong relationship between the $\delta^{18}O$ of the carbonate and its mineralogy, with more negative $\delta^{18}O$ values associated with low-Mg calcite, mixtures of low-Mg calcite, high-Mg calcite and dolomite associated with intermediate $\delta^{18}O$ values, and dolomite associated with the most positive $\delta^{18}O$ values (Fig. 5). This pattern in both $\delta^{18}O$ and mineralogy is expected in primary carbonates from a lake system undergoing varying amounts of evaporation, with high $\delta^{18}O$ values and dolomitic precipitates often associated with highly evaporated lake waters [22]. If dolomite resulted from diagenetic alteration of original calcite, we would expect much more negative $\delta^{18}O$ values and a more pervasive alteration of all the carbonate beds, not an alternating mixture of dolomite and low-Mg calcite beds.

5.2. Temperature and the $\delta^{18}O$ of surface water

Both the $\delta^{18}O$ and temperature of surface waters control the $\delta^{18}O$ of these carbonates. In lake systems, micritic carbonates typically precipitate in the late spring and early summer, as the water warms and the saturation state of the lake

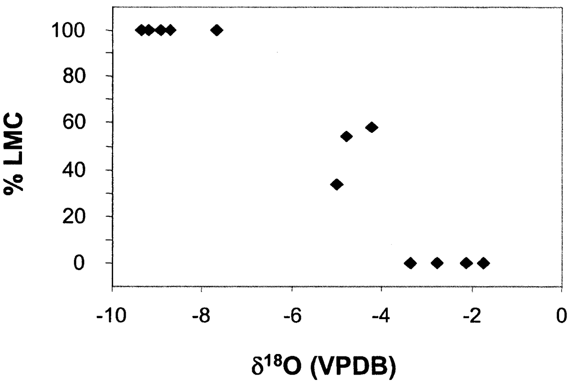

Fig. 5. Correlation of mineralogy and $\delta^{18}O$ suggests that dolomitic intervals form during times of greater evaporation and lake closure. The system is dominated by low-Mg calcite (LMC, <4 mol%) and dolomite. High-Mg calcite is present in trace amounts.

water increases [23]. This results in a relatively narrow range of formation temperatures for the lake carbonates, and the $\delta^{18}O$ of micritic carbonate is often used as a proxy for the $\delta^{18}O$ of lake water [23–26]. Fluvial and groundwater carbonates, which grow relatively slowly, would presumably be precipitated close to mean annual temperature. Note, however, that the $\delta^{18}O$ of carbonate is not strongly affected by relatively large changes in temperature; a change in temperature of 5°C leads to a change of only ~1‰ in the $\delta^{18}O$ value of carbonate [27].

The oxygen isotope compositions of the lake intervals (shaded in Fig. 4) are characterized by high-frequency oscillations of 2–6‰. This is most likely due to the varying degrees of evaporation of lake water. When the lake is open and water flows through the system relatively quickly, evaporation will be very minor and the $\delta^{18}O$ of lake water should approach the $\delta^{18}O$ of the inflowing water primarily derived from rainfall in the basin. On the other hand, closure of the lake basin will increase the residence time of water in the lake and lead to evaporation of lake water and a shift to more positive $\delta^{18}O$ values. The strong co-variance between $\delta^{18}O$ and $\delta^{13}C$ in the lacustrine intervals, commonly associated with closed lakes [24], reinforces the idea that there has been significant variation in the evaporation of the lake water (Fig. 4). Evaporation of lake water is also the simplest explanation for the mineralogy of the lake carbonates. Carbonates that are more positive than −7‰ have significant amounts of dolomite and minor amounts of high-Mg calcite, which are associated with evaporative lakes [22] (Fig. 5). Samples with more negative values are low-Mg calcite. Therefore we argue that the $\delta^{18}O$ value of lake carbonate is primarily a reflection of the $\delta^{18}O$ value of lake water and the variance in the record is due to changes in the residence time of water in the lake.

6. Discussion

The $\delta^{18}O$ value of paleometeoric water can be used to interpret changes in climate, atmospheric circulation, and/or drainage patterns. The carbonate record of the least evaporated lake waters is represented by the most negative end of the oscillations between 20 and 7.5 Ma. These data show surprising stability between ~20 and 12 Ma, with an average of ~−10.5‰ (VPDB) for the most negative lake carbonates (Fig. 4). These values are roughly congruent with the previous 9 Myr of fluvial carbonate. From 12 to about 5 Ma, the most negative $\delta^{18}O$ values average ~−9‰ (VPDB). After 7 Ma, the $\delta^{18}O$ values of fluvial carbonates have similar $\delta^{18}O$ values to the most negative lacustrine carbonates deposited between 12 and 7 Ma, consistent with little change in the $\delta^{18}O$ value of rainfall.

The positive 1.5‰ shift that occurred at 12 Ma is intriguing because of its stability over long time periods before and after this event, which suggests that the isotopic composition of rainfall remained relatively constant on either side of this shift. The abrupt change to more positive $\delta^{18}O$ values of carbonate implies a significant change in climate or in atmospheric circulation. The $\delta^{18}O$ values of rainfall in the Linxia basin can be estimated from the $\delta^{18}O$ values of the carbonates using standard calcite fractionations [27] and the typical temperature, 25±4°C, for micrite precipitation in lakes [23,24]. Prior to 12 Ma the $\delta^{18}O$ of rainfall was ~−8.5±1‰ (SMOW). At 12 Ma the $\delta^{18}O$ of rainfall changed to ~−7±1‰ (SMOW).

An alternative explanation for this 12 Ma event is that a change in temperature affected the $\delta^{18}O$ value of carbonate. This seems unlikely, as this would require a 7°C cooling that persisted for many millions of years following. Also making temperature unlikely as an explanation is the observation that temperature changes are usually associated with changes in the $\delta^{18}O$ of rainfall. A cooling of 7°C would lead to carbonates that are 1.5‰ more positive, but cooler atmospheric temperatures tend to drive the $\delta^{18}O$ of precipitation to more negative values [28]. These two trends can come close to canceling each other out; for example, a study of a New Zealand speleothem estimated that a 1°C increase in temperature would only result in a −0.05‰ change in speleothem carbonate [29]. In the case of lacustrine micrites, where precipitation events tend to occur at summer temperatures, the primary control on the $\delta^{18}O$ of carbonate is almost always the $\delta^{18}O$ of meteoric or lake water.

Another possible factor contributing to this 1.5‰ shift is a change in the $\delta^{18}O$ of seawater, which is the ultimate source of meteoric water in the Linxia basin. During the middle Miocene the $\delta^{18}O$ of Pacific Ocean benthic foraminifera increased by approximately 1.2‰ (VPDB), although it is not clear to what degree this was due to an increase in the $\delta^{18}O$ of seawater or due to a cooling of bottom waters [30]. A high-resolution stable isotope study of middle Miocene foraminifera in the South Pacific by Flower and Kennett [30] concludes that between 13.2 and 13 Ma the oxygen isotope ratios of both benthic and planktonic foraminifera increased by 0.3‰ in the Pacific, which is attributed to an increase in the $\delta^{18}O$ value of seawater [30,31][1]. They also suggest that a larger shift of 0.7‰ in the $\delta^{18}O$ of seawater occurred between 14.1 and 14.7 Ma; this change clearly preceded the increase seen in our record (Fig. 6). Therefore, if an increase in the $\delta^{18}O$ value of the seawater source occurred between 13 and 12 Ma, it made only a small contribution to the 1.5‰ change.

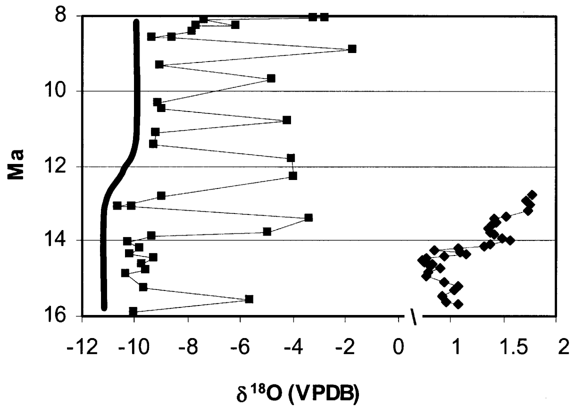

Fig. 6. The Linxia basin $\delta^{18}O$ record (left) compared to the high-resolution mid-Miocene benthic foraminifera record of Flower and Kennett [30]. The change in $\delta^{18}O$ of the 'least evaporated' lacustrine carbonates (paralleled by heavy line, sometime between 13 and 11.5 Ma) occurs later than the major change in the benthic carbonate record (which occurs prior to 14 Ma). Note expanded scale for foraminifera $\delta^{18}O$. The Flower and Kennett time scale has been modified to match that of Cande and Kent [31][1].

There are a number of other possible occurrences that could have caused this shift, but they seem unlikely in comparison to a change in atmospheric circulation and/or aridification. In this tectonically active region it is possible that a reorganization of the drainages feeding this lake system may have changed the average $\delta^{18}O$ of water coming into the lake without any major climate change. This could be tested by a comparison of $\delta^{18}O$ records from other nearby basins with that of the Linxia basin. If a similar shift occurs in an adjacent basin, then this is probably a regional climatic response to some event. Elevation changes in the region surrounding the basin could also cause a change in the $\delta^{18}O$ of runoff. Our data would imply a significant reduction in elevation in the middle Miocene, which seems unlikely. Finally, a change in the seasonal distribution of precipitation could change the average $\delta^{18}O$ of rainfall in the region. The long persistence of the change at 12 Ma argues against this, for this shift would have to persist over many millions of years, and there is no indication of any short-term return to the older −10.5‰ $\delta^{18}O$ value.

This shift to more positive $\delta^{18}O$ values is, there-

[1] Dates based on paleomagnetic chronology from papers published prior to 1995 are converted to the Cande and Kent [31] paleomagnetic time scale.

fore, most likely due to either a reorganization of atmospheric circulation patterns or a permanent change in climate. Today this region is on the edge of a major gradient in aridity, with very wet climates lying to the southeast and the arid core of China and Mongolia lying to the west. The Linxia basin receives 51 cm of rainfall per year. The path of the westerlies is blocked by the high Tibetan plateau and Pacific and Indian Ocean moisture only rarely penetrates into the region north of the plateau today. The $\delta^{18}O$ value of meteoric water in the Linxia and Lanzhou region is anomalously more positive than that of regions east and south of these cities [32] because this is a transitional zone to the very dry region immediately north of the Tibetan plateau. Regional aridity seems to shift rainfall to $\delta^{18}O$ values that are somewhat more positive than the areas supplying moisture to this region [32].

A change from −8.5 to −7‰ in meteoric water is consistent with the interruption of an Indian or Pacific Ocean source of water vapor in the Linxia region. Today the region is arid because the high Himalaya and the Tibetan plateau obstruct air masses and water vapor does not cross Tibet into central China. The $\delta^{18}O$ of rainfall in Lanzhou is −6.2‰ (VSMOW) (weighted mean) [32] and that of the Linxia basin is −6 to −7‰ (VSMOW) [33]. This agrees reasonably well with $\delta^{18}O$ values calculated for the post-12 Ma interval. The long stability of the $\delta^{18}O$ values before and after 12 Ma point to a change that is more permanent than regional climate fluctuations. The $\delta^{18}O$ value of rainfall inferred from the pre-13 Ma lacustrine carbonate (−8.5‰ SMOW) is more negative than the post-13 Ma values. More negative values are found today in locations to the east and south of Linxia, where Indian and/or Pacific Ocean moisture make a significant contribution to the annual rainfall: −10.4‰ (VSMOW) in Changqing and Kunming, −8.4‰ in Zunyi, −8.3‰ in Guiyang, and −7.5‰ in Chengdu (weighted mean) [32] (Fig. 1). After a shift to more positive values at ∼13 Ma this new isotopic ratio persisted, although there was high variability in the Plio–Pleistocene. We therefore suggest that 13–12 Ma is the point at which the Tibetan plateau, or some portion of it, rose high enough to exclude southern and/or eastern moisture from central China.

The most evaporative interval in our record occurs between 9.6 and 8.2 Ma, where the occurrence of carbonate layers in lake sediments is more frequent and they achieve the most positive $\delta^{18}O$ values. This coincides with a major change in the $\delta^{18}O$ values of soil carbonates in the Pakistan Siwalik record, which occurred at 9.3 Ma [3] (dating converted to the Cande and Kent paleomagnetic time scale [31][1]). This $\delta^{18}O$ change in Pakistan is followed by a $\delta^{13}C$ shift about 2 Myr later that reflects a shift from C3 to C4 plants. The carbon isotope record for the Linxia basin reflects aquatic processes (lacustrine, fluvial or groundwater) during this interval. It is therefore not surprising that the $\delta^{13}C$ record is very different from that of the Pakistan or Nepal soil-carbonate record [3,12]. The arid event in the Linxia basin lake record precedes a strong, but temporary, event recorded in North Pacific sediments. From 8 to 7.5 Ma a rapid increase in the aeolian dust flux occurred, achieving values that were similar to modern conditions [16]. While this increase in aridity is also apparent in our record, we note that the $\delta^{18}O$ value of meteoric water, represented by the most negative $\delta^{18}O$ values in lacustrine carbonates, does not change. This suggests that the circulation patterns and water vapor sources to the Linxia basin were similar before and after this interval. Like the Pacific dust record, this arid event was temporary. If additional uplift of the Tibetan plateau was responsible for the changes seen at ∼9 Ma in many records, it did not permanently change the source and/or composition of atmospheric vapor to the Linxia basin.

Several studies support our inference that the Tibetan plateau had already attained significant elevation by 12 Ma, sufficient to cause changes in atmospheric circulation in northeastern Tibet. Direct evidence of high surface elevation comes from the oxygen isotope composition of carbonates deposited in the Thakkhola graben beginning as early as ∼11 Ma in the southern part of the Tibetan plateau [10]. Likewise, paleoelevation estimates based on leaf physiognomy data suggest that little or no elevation change has occurred in the southern Tibetan plateau throughout the last 15 Myr [34]. In addition, a minimum age of 13.5 Ma has been obtained for an east–west extensional graben in north-central Tibet [35]. The timing of initial east–west extension has been attributed to the attainment of a threshold elevation, which led to gravitational collapse of the plateau [36–38]. The recent synthesis by Tapponnier et al. has emphasized these and many more examples of early uplift for more southerly portions of the Tibetan plateau [39], although they also suggest that uplift in the northeast of the plateau is Pliocene or later. While our data cannot be used to support or refute their three-stage uplift model of the Tibetan plateau [39], it suggests that plateau elevations were not high enough to block Indian Ocean and/or Pacific Ocean moisture from the Linxia region until 13–12 Ma. After 12 Ma some portion of the plateau was high enough to block southerly moisture.

Several climate indicators also suggest that monsoon circulation was established well before 9 Ma, including evidence of strong seasonal aridity at 10.7 Ma in the Siwaliks of Nepal [7]. Although the foraminifera [4] and diatom [40] records of the northern Indian Ocean suggest that monsoonal circulation greatly intensified at 9.3 Ma, they also suggest that monsoonal circulation was present by 10–11 Ma. As evidence mounts to support a high-elevation Tibetan plateau and intense Asian monsoon as early as 13.5–11 Ma, an alternative explanation must be explored for the cause of the 9–8 Ma event.

After the hiatus from 4.3 to 3.6 Ma there is more variability in the $\delta^{18}O$ value of fluvial carbonates than in other fluvial intervals in the basin. This is perhaps due to greater climate variability in the Plio–Pleistocene and/or due to changes in local topography and subsequent regional climate effects.

7. Conclusions

1. The high-frequency variation in the lacustrine $\delta^{18}O$ record can be explained by variation in the residence time of water in the lake system. The systematic return to a consistent $\delta^{18}O$ value at the most negative end of the range suggests that this $\delta^{18}O$ value is that of an open lake system where evaporation is minimal. This value can therefore be used to calculate the $\delta^{18}O$ of meteoric water.
2. The stable isotope compositions of fluvial and lacustrine carbonates from the Linxia basin show that between 13 and 12 Ma the $\delta^{18}O$ of meteoric water became more positive. This change was persistent and is the most significant feature in the Linxia basin record between 29 and 4.6 Ma. This is most likely a result of uplift of some portion of the Tibetan plateau, which reached an elevation that blocked Pacific or Indian Ocean moisture from reaching the region northeast of the plateau.
3. Regional aridity reaches a maximum during the interval from 9.6 to 8.2 Ma, as has been seen in many other climate records in this region. This event is a temporary feature in the record, like in the North Pacific dust flux record.

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