Early–middle Miocene topographic growth of the northern Tibetan Plateau: stable isotope and sedimentation evidence from the southwestern Qaidam basin

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Abstract

This study presents facies analysis and carbonate stable isotopic records of a 4,435-m-thick Oligocene–middle Miocene (~30 to ~11 Ma) stratigraphic section from the southwestern Qaidam basin, which place constraints on the interplay between topographic growth and climatic/hydrologic changes in the northern Tibetan Plateau. Three types of carbonate, including marl/limestone, pedogenic carbonate and carbonate cement, were collected and went through detailed screening for diagenesis. Carbonate cements are primary early diagenetic carbonates formed in shallow ground water setting. The majority of marls/limestones and pedogenic carbonates are primary micritic carbonates, whereas a few of them showing alterations were sub-sampled to avoid diagenetic phases. A negative shift in the $\delta^{18}\text{O}$ values of marls/limestones and depositional environment change from palustrine to shallow lacustrine associated with

paleohydrological change at ~20 Ma are likely caused by tectonic activity in adjacent ranges. Around 15–14 Ma, a decrease of ~1.5‰ in the most negative $\delta^{18}\text{O}$ values of carbonate cements is interpreted to reflect the topographic growth of surrounding mountain ranges (i.e., Altun Shan, East Kunlun Range). This inference is supported by the sedimentary facies change from marginal lacustrine to fluvial that was accompanied by an abrupt increase in sedimentation rate. Finally, a ~1.5‰ positive shift in the most negative $\delta^{18}\text{O}$ values of carbonate cements and pedogenic carbonates, enriched $\delta^{13}\text{C}$ values of pedogenic carbonates, as well as sedimentary environment change from fluvial to braided river/alluvial fan at 13–12 Ma, most likely reflect a regional aridification event observed over a large area of the Central Asia. This aridification event may have been associated with continued topographic growth of the northern Tibetan Plateau to a critical elevation to block moisture (e.g., Westerlies) from penetrating the interior of Central Asia.

Keywords: Tibetan Plateau; Qaidam basin; Stable isotope; Facies analysis; Topographic growth; Paleoclimate

1 Introduction

Numerical climate simulations have shown that the topographic growth of the Tibetan Plateau exerts significant influence on the climate of Asia (Liu and Yin, 2002; Prell and Kutzbach, 1992; Roe et al., 2016; Ruddiman and Kutzbach, 1989; Zhang et al., 2007b), resulting in the intensification of Asian summer monsoons in the southern and eastern Tibetan Plateau, as well as aridification in the northern and western Plateau (An et al., 2001). The topographic growth itself and associated climatic changes (e.g., changes in moisture sources, rain shadow/moisture blocking effects, enhanced sub-cloud evaporation and/or surface water recycling; Blisniuk and Stern, 2005) influence the stable isotopic values of precipitation and hence soil, lake and fluvial waters, as well as authigenic carbonates that precipitate from those waters. In addition to topographic growth, retreat of the Para-Tethys seaway from the Tarim basin (Bosboom et al., 2014; Harzhauser and Piller, 2007; Rögl, 1998) would greatly influence the source of moisture to the interior of Central Asia (Fluteau et al., 1999; Ramstein et al., 1997; Zhang et al., 2007a). Likewise, Cenozoic global cooling would result in a temperature decrease and a change toward more arid condition (Dupont-Nivet et al., 2007; Ge et al., 2013; Kraatz and Geisler, 2010; Lu et al., 2010; Miao et al., 2012). These factors could also result in isotopic changes in authigenic carbonates and should be carefully evaluated.

Multiple studies have used the stable isotopic variations of sedimentary carbonates to reveal the paleoclimate and topographic evolution of the northern Tibetan Plateau (Dettman et al., 2003; Fan et al., 2007; Graham et al., 2005; Hough et al., 2014; Hough et al., 2011; Kent-Corson et al., 2009; Rieser et al., 2009; Zhuang et al., 2011a). Most of the studies infer topographic growth of surrounding ranges since 15–14 Ma (Hough et al.,

2014; Hough et al., 2011; Zhuang et al., 2014) or later (e.g., 12 Ma; Dettman et al., 2003; Zhuang et al., 2011a). However, evidence from other proxies (e.g., low temperature thermochronology, magnetostratigraphy) suggests there were active tectonic phases associated with exhumation in the northern Plateau during the latest Oligocene–earliest Miocene (e.g., 24–20 Ma; Hough et al., 2014; Lease et al., 2012; Sobel and Dumitru, 1997; Wang et al., 2003; Yuan et al., 2006), as well as much earlier in the middle Eocene (Clark et al., 2010; Duvall et al., 2013; Yin et al., 2008). Kent-Corson et al. (2009) reported negative $\delta^{18}\text{O}$ shifts of carbonate rocks in several sedimentary sections in the Qaidam and Tarim basins during the Paleogene and attributed these to topographic growth of the northern Tibetan Plateau although the timing was loosely confined.

Qaidam basin is the largest Cenozoic basin in the northern Tibetan Plateau (Figure 1B) and records abundant regional tectonic information from Paleogene to Neogene time (Zhuang et al., 2011b). Previous stable isotopic studies in the Qaidam basin either reflect a short temporal range (e.g., <15 Ma; Zhuang et al., 2011a), or limited age constraints (Graham et al., 2005; Kent-Corson et al., 2009), or poor sampling resolution (Rieser et al., 2009). This study presents new stable isotopic data from a 4,435-m-thick sedimentary section in the southwestern Qaidam Basin (Figure 1C). With age constraints that extend from the Oligocene to middle Miocene (~31 to ~11 Ma; Chang et al., 2015), as well as detailed sedimentary facies analysis, we provide new insights into the climatic/hydrologic and topographic evolution of the northern Tibetan Plateau.

2 Geological setting, stratigraphy and age constraints

The studied Huatugou section lies in the southwestern Qaidam basin (Figure 1B), which is a large (~120,000 km²) intermontane basin located near the northern edge of the

Tibetan Plateau (Figure 1A). With elevations between 2,600–3,000 m, the Qaidam basin is surrounded on all sides by 4,000–6,000 m high mountain ranges, which are also bounded by large fault systems. The basin is bounded by the Altun Shan (“Shan” in Chinese means mountain) and Altyn Tagh fault to the northwest, the East Kunlun Range and Kunlun fault/Qimen Tagh thrusts to the south, and the Qilian Shan and north Qaidam thrusts to the northeast (Figure 1B). Tectonic activity in the mountain ranges and fault systems of the Altyn Tagh and Qimen Tagh provided the majority of detritus and controlled sedimentary evolution in the western Qaidam basin (Cheng et al., 2015a; Rieser et al., 2005; Wang et al., 2006).

The depositional center of the Qaidam basin initially lay close to the Altyn Tagh fault and then migrated to the east, probably due to the uplift of the Altun Shan and intra-basinal deformation (Meng and Fang, 2008; Yin et al., 2008). Deformation within the Qaidam basin, manifested as northwest–southeast trending folds, such as the Youshashan anticline (Figure 1C), initiated during the late Miocene, and continued through the Quaternary (Meng and Fang, 2008; Yin et al., 2008).

Cenozoic strata with a total thickness of ~6–7 km have filled the Qaidam basin (Huang et al., 1996). The Cenozoic strata, composed mainly of terrestrial alluvial-fluvial to lacustrine deposits, have been divided into seven formations, the ages of which were constrained by magnetostratigraphy (Chang et al., 2015; Fang et al., 2007; Lu and Xiong, 2009; Song et al., 2013; Sun et al., 2005b; Zhang, 2006; Zhang et al., 2014). From old to young, these seven formations are: Lulehe (early–middle Eocene), Xiaganchaigou (middle–late Eocene), Shangganchaigou (Oligocene), Xiayoushashan (early–middle Miocene), Shangyoushashan (middle–late Miocene), Shizigou (late Miocene–Pliocene)

and Qigequan (Pleistocene).

Using magnetostratigraphy, Chang et al. (2015) determined ages for most of the Huatugou section (Figure 2), which has a total stratigraphic thickness of 4,435 m. They correlated strata between 1,200–4,100 m to 20.7–11.6 Ma of Geomagnetic Polarity Time Scale (GPTS; Gradstein et al., 2012). For the lower part of the section between 0–1,200 m, however, they suggested two possible correlations. One designates ~28–20.7 Ma, and the other suggests an age range of ~31–20.7 Ma (Figure 2). This second correlation accords with another magnetostratigraphic study in the same area (within hundreds of meters distance away), which correlates to ~31 Ma for the bottom of the Cenozoic strata in the Huatugou area (Zhang, 2006). Regardless of which correlation for the lower part is used, there is no impact on the ages for the isotopic changes that we discuss, because all the significant isotopic shifts are above 1,200 m.

3 Methods

3.1 Stratigraphic study

The Huatugou section crops out along the steep southern limb of the Youshashan anticline in the southwestern Qaidam basin (Figure 1C). This section includes the upper part of the Xiaganchaigou (0–470 m, Figure 2), the Shangganchaigou (470–1,940 m), Xiayoushashan (1,940–3,100 m) and Shangyoushashan (3,100–4,360 m) formations. The uppermost ~80 m of thick conglomerate (4,360–4,435 m) belongs to the Shizigou Formation. Detailed lithology, primary sedimentary structure, grain size, bedding thickness, bedding boundaries and lateral extent were recorded in the field.

3.2 Sample collection

In addition to lacustrine limestone/marl and pedogenic carbonate, carbonate cement

was also collected. Carbonate cement can be faithful paleoclimate and paleoenvironment recorder (Alonso-Zarza, 2003; Mack et al., 2000; Quade and Roe, 1999), if it was formed during early diagenesis, such as within the vadose zone or near the shallow ground water table (Choquette and Pray, 1970; Morad, 1998). In total, we collected 128 authigenic carbonates, including 17 pedogenic carbonate nodules, 57 lacustrine limestones and marls, and 54 carbonate cements (Table 1).

3.3 Petrographic study

Twenty out of the 128 carbonate samples were cut into thin sections for petrographic study under microscope. The twenty thin section samples include 4 marls, 4 limestones, 5 pedogenic carbonates and 7 carbonate cements, which cover all types of carbonate and the range of stratigraphic levels in the section, as well as the most positive, most negative, and intermediate oxygen isotopic values of all types of carbonate (Table 1). Under the microscope, the following features were examined: the texture (micrite, microspar, or sparite), the characteristics of porosity (primary or secondary), the percentage of cement by volume, the mineralogy (e.g., detrital carbonate grains), the presence and characteristics of vugs and infilling spar, and the presence and characteristics of vein calcite.

3.4 Laboratory analysis

A small piece of each sample that shows no visible veins or recrystallization was crushed to fine powder. Then, 1 ml 30% H₂O₂ was added to react with the powder for 20 minutes to remove organic matter; after which the powder was rinsed with deionized water then dried at room temperature. The dried powder was weighed and sealed in a glass vial, flushed with helium gas, and then reacted with pure phosphoric acid at 70°C to

release carbon dioxide gas. Analyses were carried out using a Thermo Delta plus XP CF-IRMS with a GasBench II peripheral device in the University of Rochester SIREAL laboratory. Isotopic results for oxygen and carbon are reported using standard delta notation ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) with respect to Vienna Pee Dee Belemnite (VPDB). Three in-house standards calibrated to the international standards NBS-18, NBS-19, and L-SVEC were used to calculate isotopic ratios. Analytical errors for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ are smaller than $\pm 0.1\text{‰}$ (1σ) and $\pm 0.06\text{‰}$ (1σ), respectively.

4 Facies analysis

The lowermost part (0–200 m, Figure 2) of the Huatugou section consists mainly of matrix- and clast-supported gravel to pebble sized conglomerate. Conglomerate is massive, poorly sorted, sub-rounded, and occurs in <1 m thick lenticular beds that stack to form 1–10 m thick conglomerate lenses. Beds show erosive bases and limited lateral extent (several 10s of meters). Massive and/or horizontally stratified sandstone and mudstone layers, 0.5–2 m thick, are interbedded with conglomerate. We interpret these sediments to be deposited in a braided river setting.

From 200 to 2,700 m (Figure 2), deposits are dominated by thinly bedded fine- to medium-grained, massive and/or horizontally stratified sandstone and mudstone. Lenticular multistory conglomerate bodies are interbedded in sandstone layers, and are 1–10 m thick and 10s–100s m in lateral extent. Interbedded, thinly bedded limestone and marl (Figure 3A) are present throughout this part of the section. Sandstone sometimes shows structures related to soft-sediment deformation. Varved deposits with fossil snails, moth heads, and feathers, as well as sandstone with rippled cross laminations were also observed. These features suggest that this part of the section is dominated by a marginal

lacustrine environment. Two sub-environments within this marginal lacustrine setting are distinguishable. Between 200 to 1,400 m, abundant gypsum (1–3 cm thick) and paleosol layers (Figure 3B) indicate relatively long periods of lake drying and subaerial exposure in arid climate intervals. Ripple cross-laminated sandstones that drape gypsum deposits record subsequent water inundation. We infer that these deposits were accumulated in lake margin shallow lacustrine to palustrine environments. Between 1,400 and 2,700 m, limestone and marl become more abundant components of lacustrine deposits, whereas gypsum layers were not observed. Thus, although generally still interpreted as a marginal lacustrine setting, there was some open lacustrine deposition during this time interval.

The 2,700–3,500 m part of the Huatugou section (Figure 2) is also dominated by fine-grained deposits, including massive mudstone and fine-grained sandstone with traction flow depositional structures, such as trough and planar cross stratifications, as well as scour features. Compared with the previous part (200–2,700 m), no marl or limestone is present, whereas sandstone and conglomerate layers representing channel deposits are more abundant. Conglomerate beds (~0.5 to 2 m thick) of limited lateral extent (10s of meters) are intercalated as lenses. Paleocurrent reconstructions from 72 imbricated clasts at 7 different stratigraphic heights (~10 clasts per site) between 2,700–3,000 m indicate that sediments were transported from the west (mean paleoflow direction = 92°, Figure 2). Carbonate concretions (carbonate cements), also known as calcrete (Wright, 2007), were frequently observed in sandstone. These carbonate concretions are generally elongated in shape and sub-parallel to the bedding surface, which denotes shallow ground water influence (Figure 3C). Pedogenic carbonate nodules are also more frequent than in the previous part, and were formed in clay- to silt-rich

201 massive mudstones in the lower part of the B horizon of paleosols, which were identified
202 as type II paleosol (Gile et al., 1966) or gleysol to calcisol (Mack et al., 1993). We infer
203 that deposition of this part of the section occurred in a fluvial environment with extensive
204 coarse channel deposits and fine-grained floodplain deposits.

205 Compared with the underlying sediments (2,700–3,500 m), the upper part of the
206 Huatugou section (3,500–4,435 m) is characterized by more prevalent coarse sandstone
207 and conglomerate layers (Figure 2) that are up to several meters thick, and tens of meters
208 in lateral extent. Although less prevalent than lower in the section, fine-grained
209 sandstones, siltstones and mudstones that extend at least 100 meters laterally (as far as
210 they can be traced) denote more limited floodplain environments. We interpret these as
211 channel and floodplain deposits of braided river system, possibly within a distal alluvial
212 fan depositional environment (Miall, 1977). Paleosols were poorly developed, but root
213 traces are common. This indicates short exposure time for soil development, probably
214 due to frequent channel migration on the alluvial fan. Deposits coarsen upward toward
215 the top of the section, where there is ~100 m thick coarse conglomerate, suggesting
216 progradation of the alluvial fan to the Huatugou area. Paleocurrent reconstruction of
217 paleoflow directions show a change from eastward flow (92°) to southwestward flow
218 (220°, mean direction based on 20 imbricated clasts at 2 sites around 3,700 m; Figure 2),
219 which suggests a shift of sediment source to local highlands northeast of the study area
220 (Figure 1B). Growth strata observed in this part supports a late Miocene deformation of
221 the Youshashan anticline (e.g., Meng and Fang, 2008; Yin et al., 2008).

5 Stable isotope controlling factors and diagenesis assessment of carbonates

5.1 Stable isotopes of primary carbonates

There are three major types of primary carbonates in terrestrial depositional environment: lacustrine (lake) limestone/marl, pedogenic (soil) carbonate and shallow ground water carbonate cement (Figure 4). Due to the fact that different primary carbonates are formed in different hydrological and depositional environments, their $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values are controlled by different factors, which are summarized in Table 2. We focus our discussions on carbonate oxygen isotopic ratios as they are tied more strongly to climate and environment.

The oxygen isotopic composition of primary carbonates are largely controlled by the isotopic signature of surface water from which they precipitate (e.g., lake water, soil water, and ground water) as well as the temperature of carbonate precipitation (Table 2, Figure 4). A large change (5°C) in ambient temperature only leads to a small change in oxygen isotopic value ($\sim 1\text{‰}$) of carbonate (Kim and O'Neil, 1997). Rainfall and snowfall are the primary sources of surface waters. The isotopic value of precipitation in one location is influenced by the local air temperature at which water condenses from atmospheric vapor, as well as the source of moisture and atmospheric circulation (Table 2). Precipitation that infiltrates into the shallow surface becomes soil water. Evaporation of soil water is insignificant in climates that receive >300 mm of rainfall per year (Quade et al., 2007); thus, the $\delta^{18}\text{O}$ value of soil water resembles that of precipitation. Likewise, shallow ground water also originates from infiltration of precipitation (Figure 4), and thus reflects the isotopic signal of precipitation, assuming minimal evaporation (Table 2). When unconsolidated clastic sediments become cemented by carbonate under near

surface conditions, carbonate cement will record the $\delta^{18}\text{O}$ value of shallow ground water (Mack et al., 2000; Quade and Roe, 1999).

River water and ground water converge into lowland settings to become lake water, where lacustrine limestone and marl are deposited (Figure 4). Compared with soil water and ground water, the oxygen isotopic composition of lake water in low relief settings tends to be enriched in ^{18}O due to evaporation. However, lake basins that are surrounded by high relief topography may collect precipitation from high elevation drainages, and therefore water in the lake may have more negative $\delta^{18}\text{O}$ value than local low-elevation precipitation. In addition, the $\delta^{18}\text{O}$ value of lake water can also change associated with hydrological changes from open lake conditions (i.e., significant outlet stream) to closed lake conditions (i.e., minimal to no stream outlet). Carbonates deposited in evaporating closed-basin lakes usually show positive isotopic excursions and covariance between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values (Talbot, 1990).

Primary carbonates are susceptible to diagenesis after deposition, which can modify the original isotopic compositions. In general, carbonate diagenesis that occurs at burial temperatures that are higher than surface temperatures will result in more negative $\delta^{18}\text{O}$ values of carbonate if carbonate recrystallization or precipitation occurred in the presence of water of similar $\delta^{18}\text{O}$ composition to surface waters (Garzzone et al., 2004). It is therefore essential to evaluate the diagenetic history of the carbonate rocks before making any paleoclimate interpretations.

5.2 Screening for diagenesis of carbonate cements

Several lines of evidence suggest that the carbonate cements above 500 m in the Huatugou section are primary carbonates that were formed during early diagenesis, and

270 did not experience significant burial diagenetic alteration. First, for all the carbonate
271 cement samples examined under microscope (Table 1), the volume percentage of cement
272 is between 30% and 40% (e.g., Figure 5A–D), which is a typical range of primary
273 porosity at shallow depth for sandstones and siltstones. Second, the preservation of
274 unaltered feldspar and mica minerals (Figure 5C) suggests that the samples were not
275 compacted by deep burial (i.e. >5 km), as these minerals tend to be altered to clay
276 minerals under deep burial diagenesis (Fan et al., 2014; Morad, 1998). Third,
277 petrographic observations show the cements are dominated by micrite and local
278 microspar (Figure 5A–D). Fourth, vugs with near-equidimensional shapes are common
279 and lack calcite infillings and surrounding calcite recrystallization (Figure 5B–D). Fifth,
280 veins and vein calcite denoting deep burial fluid activity are rarely present, both in thin
281 sections and hand samples. These observations suggest that carbonate cements are
282 primary carbonates, formed during early diagenesis. We further suggest that they were
283 formed in shallow ground water environment (Figure 4), based on the observation that
284 the carbonate cements in the field are expressed as elongated concretions and layered
285 calcretes that are parallel to bedding (Figure 3C). This unique shape and occurrence
286 strongly suggest that they were formed at the interface between the vadose zone and the
287 shallow ground water table, and thus record the isotopic signals that reflect surface
288 conditions.

289 We cannot rule out the possibility that carbonate cements with more negative $\delta^{18}\text{O}$
290 values ($\sim -10\text{‰}$) at the bottom of section (<500 m, Figure 6) might have undergone deep
291 burial diagenesis and therefore do not record the isotopic composition of surface waters
292 in the context of climate change and/or changes in the hydrology of the basin.

5.3 Screening for diagenesis of pedogenic carbonate, marl and limestone

Petrographic observations show that most of the pedogenic carbonates, marls and limestones retain their original micritic fabric (Figure 5 E–G), indicating minimal diagenetic alteration. The $\delta^{18}\text{O}$ values of limestones, marls and pedogenic carbonates are consistent with the oxygen isotopic values of surface waters from which they precipitate (e.g., 11HTG45, limestone, -1.2‰ ; 11HTG96, marl, -6.1‰ ; and 11HTG124, pedogenic carbonate, -8.5‰ ; Table 1), reflecting the degree of evaporative enrichment of ^{18}O in each depositional environment (Figure 4). The lack of increasingly negative $\delta^{18}\text{O}$ values of marls, limestones and pedogenic carbonates toward the lower part of section suggests that burial did not alter the original isotopic compositions of carbonates.

There are a small number of limestones that show signs of diagenetic alteration, where the altered regions are characterized by small (<1 mm) clusters of blocky sparry calcite or microspar around vugs (Figure 5 H–I); the latter of which suggests alteration was focused around vugs. During sample crushing for isotopic analysis, partially altered samples were sub-sampled using a micro-drill to avoid diagenetic phases.

6 Stable isotope results and interpretations

We break the Huatugou isotopic section (500–4,435 m) into four phases based on the variations in the $\delta^{18}\text{O}$ values (Figure 6, see Table 1 for all the isotopic data). The lower 500 m was not considered as this part may have experienced diagenetic alteration (see section 5.2).

From 500 to 1,300–1,400 m (phase I; 27.8 or 24.7 Ma to ~ 20 Ma), the $\delta^{18}\text{O}$ values of limestones and marls show large variability between -1.2‰ and -7.4‰ , whereas carbonate cements tend to have more negative values between -7.2‰ and -9.0‰ . From

1,300–1,400 m to 2,600–2,700 m (phase II; ~20 Ma to 15–14 Ma), the $\delta^{18}\text{O}$ values show a more narrow range of lower values between -5.9‰ and -8.9‰. Although positive-negative excursions (marl/limestone vs. carbonate cement) still exist, they are significantly damped to within 3.2‰ variation, compared with the ~8‰ variation in phase I. Although the $\delta^{18}\text{O}$ values of carbonate cements still mark the lower limit of marls/limestones and remain unchanged throughout this interval, unlike phase I, they are no more negative than the most negative marls/limestones. Above 2,600–2,700 m and below ~3,600 m (phase III; 15–14 Ma to 13–12 Ma), the most negative $\delta^{18}\text{O}$ values of carbonate cements decrease from -8.7‰ of phase II to ~-10.0‰. Except three samples, the $\delta^{18}\text{O}$ values of all the other carbonate cements and pedogenic carbonates are between -10‰ and -8.5‰. Finally, in the upper part of the section (>3,600 m, phase IV; 13–12 Ma to ~11 Ma), the most negative $\delta^{18}\text{O}$ values show a positive shift from -10.0‰ to -8.6‰. All three pedogenic carbonates and five carbonate cements in this phase show values between -8.2‰ to -8.6‰.

6.1 Early Miocene (~20 Ma) hydrological change in the southwestern Qaidam basin

The most negative $\delta^{18}\text{O}$ values of carbonate cements remain the same from >500 m in Phase I to ~2,600 m of Phase II (Figure 6, green thick line), suggesting that the isotopic composition of local precipitation did not change, and thus no significant regional climate change occurred at ~20 Ma (Figure 7A and 7B). Compared with carbonate cements, there are large negative shifts of the $\delta^{18}\text{O}$ values for marls/limestones: ~5‰ for the most positive values (Figure 6, black solid curve); and ~2‰ for the most negative values (Figure 6, black dotted curve). The most positive $\delta^{18}\text{O}$ values of marls/limestones are due to evaporative enrichment of lake water (Figure 4), whereas the

most negative values reflect the least evaporative enrichment, and thus are more close to the isotopic composition of precipitation.

Since the unchanging $\delta^{18}\text{O}$ values of carbonate cements do not support a major climatic change, we suggest that the negative $\delta^{18}\text{O}$ shift of marls/limestones from phase I to phase II reflect a paleo-hydrological change in the southwestern Qaidam basin around 20 Ma: from a restricted sub-basin (Figure 7A) to an open basin (margin of a much larger lake basin; Figure 7B). In restricted water body, such as a sub-basin, the smaller water volume and restricted connection with the remaining larger water body make it more prone to variations in water composition associated with evaporative enrichment, which would result in more positive oxygen isotopic values of marls and limestones. By contrast, when the sub-basin is incorporated into a much larger lake basin, the buffering effect of the large lake volume would suppress extreme evaporative enrichment of ^{18}O . This would result in smaller positive deviations from local meteoric water (such as precipitation, river water and ground water), which would manifest as overall more negative $\delta^{18}\text{O}$ values that show a smaller range between the most negative and most positive values.

This hydrological change deduced from isotopic data is supported by our facies analysis. Between 200 and 1,400 m, the depositional environment is dominated by palustrine deposits with abundant gypsum layers, whereas between 1,400 m and 2,700 m, open lacustrine deposits dominate: marls/limestones become more important components, while gypsum layers are absent (Figure 2). Depositional thickness contours (Meng and Fang, 2008) provide additional evidence for the hydrological change. During and before the Oligocene, a sub-basin existed in the southwestern Qaidam basin (Figure 7A); this sub-basin disappeared during the early Miocene, and was incorporated into a larger lake

environment (Figure 7B).

6.2 Early middle Miocene (15–14 Ma) range growth in the northern Tibetan Plateau

Around 2,600–2,700 m (15–14 Ma), the most negative $\delta^{18}\text{O}$ values of carbonate cements decrease from $\sim -8.5\%$ of phase II to about -10.0% of phase III (Figure 6, green thick line). We interpret this isotopic change as a consequence of isotopic change of precipitation. The isotopic similarity between carbonate cements and pedogenic carbonates confirms that the isotopic values of carbonate cements reflect those of precipitation (Figure 4). In addition to isotopic change, we also observed sedimentary facies change from marginal lacustrine to fluvial channel-floodplain setting (Figure 2). Magnetostratigraphic study of this section also found an abrupt increase in the sedimentation rate at ~ 15 Ma from ~ 12.5 cm/kyr to ~ 50 cm/kyr (Chang et al., 2015). A negative oxygen isotopic shift was also observed in several other sedimentary sections that lie on the margins of the northern Tibetan Plateau (Kent-Corson et al., 2009), although their timing of transitions are less precisely resolved than the Huatugou section described here.

One possible explanation for this negative oxygen isotopic shift and depositional change is tectonic activity and growth of surrounding ranges (Figure 7C). An elevated mean elevation of catchment regions (e.g., Altun Shan, Qimen Tagh Range) would lead to more negative oxygen isotopic values of precipitation falling in this region, and correspondingly more negative $\delta^{18}\text{O}$ values of carbonates (e.g., Graham et al., 2005; Kent-Corson et al., 2009). Increased relief would also lead to increased erosion rate, and thus increased sedimentation rate, perhaps resulting in progradation and filling of the lacustrine basin and a facies transition from shallow lacustrine to fluvial environment

(Figure 7C).

Global cooling of $\sim 3^{\circ}\text{C}$ at ca. 14 Ma (e.g., Flower and Kennett, 1993; Zachos et al., 2001) is unlikely to be the only cause of the 1.5‰ decrease in $\delta^{18}\text{O}$ values at 15–14 Ma. Although a decrease of air temperature could impart a decrease in the $\delta^{18}\text{O}$ values of precipitation and surface waters, a decrease in the temperature of carbonate precipitation would increase the $\delta^{18}\text{O}$ values of carbonate, tending to offset the potential decrease in rainfall $\delta^{18}\text{O}$ values (e.g., Dettman and Lohmann, 2000).

6.3 Late middle Miocene (13–12 Ma) aridification of the Central Asia

Toward the top of the Huatugou section, there is a positive shift in the most negative oxygen isotopic values from $\sim -10.0\text{‰}$ of phase III to $\sim -8.6\text{‰}$ of phase IV around 13–12 Ma (Figure 6, green thick line). As this shift is expressed by both carbonate cements and pedogenic carbonates, which are more representative (compared with marl and limestone) of local rainfall composition (Figure 4), it likely reflects intensified aridification. The fact that the 13–12 Ma isotopic shift is observed in pedogenic carbonates that occur both above and below this boundary (Figure 6) further supports the inference that this shift was caused by climate change (e.g., intensified aridification), rather than a change in carbonate depositional environment.

Intensified aridification could increase the oxygen isotopic values of meteoric water through three processes, the first of which is increase in the percentage of recycled local surface water, including lake water and soil water. Evaporated local surface water usually has higher $\delta^{18}\text{O}$ values due to kinetic fractionation under low relative humidity (arid) conditions (Gonfiantini, 1986). Mixing of these relatively positive water with atmospheric water vapor would lead the $\delta^{18}\text{O}$ values of subsequent precipitation to be

more positive (Bershaw et al., 2012). Intensified sub-cloud evaporation is another process that can increase the $\delta^{18}\text{O}$ values of precipitation as raindrops fall through unsaturated air column between cloud base and the ground (Lee et al., 2012; Li and Garziona, in revision). Lastly, preferential removal of light isotopes by intensified evaporation can cause the $\delta^{18}\text{O}$ values of surface water (i.e., soil water) to become more positive than original rainfall composition.

The inference of an arid climate condition is further supported by the observation of very positive ($>-5\text{‰}$) $\delta^{13}\text{C}$ values of pedogenic carbonates. During the middle Miocene, C_4 plants were not present in the Qaidam basin (Wang et al., 2007). Therefore, high $\delta^{13}\text{C}$ values of pedogenic carbonates suggest lower soil respiration rates, associated with dry climate conditions (Cerling and Quade, 1993). We also observe a sedimentary environment change from fluvial to braided river and alluvial fan (Figure 2). Coarse grained alluvial fan deposits dominated by conglomerates are generally more common sedimentation types in arid climate conditions (Blair and McPherson, 2009).

Similar to the Huatugou section in the southwestern Qaidam basin, a positive shift in oxygen isotopic values around 12 Ma was observed across a large area of the northern Tibetan Plateau and surrounding regions including the Tarim basin and several smaller basins in the northeastern Plateau (Dettman et al., 2003; Hough et al., 2011; Kent-Corson et al., 2009; Zhuang et al., 2011a). This suggests that the positive shift in oxygen isotopic values reflects a regional climatic change associated with intensified aridification.

Previous studies have argued that this positive shift of oxygen isotopic values (intensified aridification) was caused by upward and outward growth of the northern Tibetan Plateau in the middle Miocene (Dettman et al., 2003; Hough et al., 2011).

Growth of the northern Tibetan Plateau, as well as the Pamir and Tian Shan, would block most moisture that was transported by Westerlies from penetrating the interior of the northern Plateau. As Westerlies provide the majority of moisture in northwest China (Araguás-Araguás et al., 1998; Li and Garzzone, in revision; Tian et al., 2007), this blocking of moisture would cause intensified aridity in the Qaidam and Tarim basins.

Global cooling is not likely to be a controlling factor for the 13–12 Ma intensified aridification as this isotopic shift postdated cooling at 14 Ma (Flower and Kennett, 1993; Zachos et al., 2001). Because the Para-Tethys sea retreated from the Tarim basin no later than the Oligocene (e.g., 37 Ma; Bosboom et al., 2014; Sun and Jiang, 2013; Wei et al., 2013), retreat of the Para-Tethys Sea is also unlikely to be a contributor for the aridification event around 12 Ma. We also do not consider a change of moisture source (e.g., from Indian summer monsoon with lower $d^{18}O$ values to Westerlies with higher $d^{18}O$ values; Tian et al., 2007) as a possible reason, considering the establishment of high topography in the southern Tibetan Plateau probably occurred long before Miocene (Ding et al., 2014; Kapp et al., 2007; Wang et al., 2014), which would have blocked moisture from the Indian summer monsoon from entering the northern Plateau (Boos and Kuang, 2010; Caves et al., 2015). It has been suggested that the northern Tibetan Plateau was dominated by Westerly moisture since the early Eocene time (Caves et al., 2015).

7 Early–middle Miocene topographic growth of the northern Tibetan Plateau

Based on the carbonate stable isotope and facies analysis evidence from the Huatugou section, we propose a topographic growth history in the northern Tibetan Plateau during the early–middle Miocene (>20 to ~11 Ma).

Between 31 and 20 Ma, the Huatugou region in the southwestern Qaidam basin

was occupied by a small sub-basin (Meng and Fang, 2008) dominated by palustrine facies (Figure 7A). Poor connectivity with the main lake basin would have made this region more prone to evaporative enrichment of ^{18}O , thus the $\delta^{18}\text{O}$ values of primary marls and limestones are much higher than those of carbonate cements (Figure 6, phase I).

At ~20 Ma, the previous sub-basin was incorporated into the larger main basin (Figure 7B). The isotopic buffer effect of the larger volume of lake water suppressed evaporative enrichment of ^{18}O , causing the $\delta^{18}\text{O}$ values of marls/limestones to be more representative of local precipitation, thus more negative (Figure 6, phase II). This paleohydrological change was probably associated with an early phase of tectonic activity/range growth along the Altyn Tagh fault (Altun Shan) and the East Kunlun Range, which has also been documented in other studies (Jolivet et al., 2001; Mock et al., 1999; Ritts et al., 2004; Wang et al., 2004; Yuan et al., 2006; Yue et al., 2001). The similarity of $\delta^{18}\text{O}$ values between carbonate cements and the most negative marls/limestones (phase II, Figure 6) suggests that only minimal topography existed in the surrounding ranges (Figure 7B).

Around 15–14 Ma, a major pulse of topographic growth of the Altun Shan and East Kunlun Range (Chang et al., 2015; Cheng et al., 2015b; Jolivet et al., 2003; Ritts et al., 2008; Yuan et al., 2013; Yue et al., 2003) caused an increase in the mean catchment elevations of the southwestern Qaidam basin (Figure 7C). This led to the negative shift of the $\delta^{18}\text{O}$ values of carbonate cements (Figure 6, phase III), sedimentary facies change from marginal lacustrine to fluvial (Figure 2), as well as abrupt increase in sedimentation rate (Chang et al., 2015).

Finally, we infer that the documented growth of more distal mountain ranges in the northern Tibetan Plateau (Craddock et al., 2011; Fang et al., 2007; George et al., 2001; Lease et al., 2012; Lu and Xiong, 2009; Sun et al., 2005a; Wang et al., 2003) and possibly also in the Tian Shan regions of central Asia (Bullen et al., 2001; Sobel and Dumitru, 1997) reached elevations sufficient to block moisture from penetrating the interior of the northern Tibetan Plateau and Central Asia by 13–12 Ma (Figure 7D), resulting in aridification in these regions, and the positive oxygen isotopic shift of precipitation and carbonates (Figure 6, phase IV).

Abundant evidence from magnetostratigraphy, sedimentary facies, sedimentation rate, stable isotopes, and low temperature thermochronology, suggests early–middle Miocene tectonic activity and topographic growth in the northern Tibetan Plateau and central Asia (see summary in Craddock et al., 2011 and Yuan et al., 2013). Our stable isotope studies and facies analysis in the southwestern Qaidam basin provide additional evidence to indicate that the topographic growth of the northern Tibetan Plateau during the early to middle Miocene time also influenced the hydrology and climate of this region (Figure 7B-D).

8 Conclusions

We carried out stable oxygen and carbon isotope studies in an Oligocene–middle Miocene (~31 to ~11 Ma) sedimentary section in the southwestern Qaidam basin (Huatugou section), northern Tibetan Plateau. We emphasize that different carbonate types (carbonate cement, marl/limestone and pedogenic carbonate) should be evaluated independently due to their diverse formation environments and isotopic fractionation mechanisms. Detailed microscopic work suggests that carbonate cements are primary

early diagenetic carbonates; the majority of marls/limestones and pedogenic carbonates are primary, and were sub-sampled to avoid diagenetic phases in the few samples that show evidence of diagenesis. Thus, the isotopic data reflect surface conditions (e.g., climatic and hydrologic changes).

Oxygen isotopic data of carbonate rocks reflect continued tectonic activity and topographic growth in the northern Tibetan Plateau during the early and middle Miocene time. The negative shift in the $\delta^{18}\text{O}$ values of marls/limestones and depositional environment change from palustrine to shallow lacustrine at ~20 Ma were probably caused by early tectonic activity along the Altyn Tagh fault and Qimen Tagh ranges that resulted in the integration of the southwestern sub-basin into the larger Qaidam lake basin. A significant pulse of range growth around 15–14 Ma is inferred based on the observation of a negative shift of the most negative $\delta^{18}\text{O}$ values of carbonate cements, sedimentary facies changes, and an increase in sedimentation rate. Continued topographic growth of the northern Tibetan Plateau and surrounding areas began to block moisture (e.g., Westerlies) from penetrating the interior of the northern Plateau and Central Asia by 13–12 Ma, leading to a positive shift in the oxygen isotopic values as supported by previous paleoclimate inferences of a regional aridification event.

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Captions

Table 1. Stable isotopic data ($d^{13}C$ and $d^{18}O$) of carbonates of the Huatugou section. Also shown are their types and stratigraphic heights.

Table 2. Controlling factors of oxygen and carbon isotopic compositions of different primary carbonates in terrestrial environments.

Figure 1. A: Topographic map of the Tibetan Plateau, with box showing the location of 1B. B: Tectonic and topographic map of the Qaidam basin and surrounding regions, showing major faults. Box shows the location of 1C. C: Geological map of studied area. Legend: Q2, Holocene; Q1, Pleistocene; N2, late Miocene–Pliocene (Shizigou Formation); N1–2, middle–late Miocene (Shangyoushashan Formation); N1–1, Early–middle Miocene (Xiayoushashan Formation); E3, Oligocene (Shangganchaigou Formation); E2, middle–late Eocene (Xiaganchaigou Formation); K, Cretaceous; J1–2, Early–Middle Jurassic.

Figure 2. Stratigraphic column of the Huatugou section with sedimentary environment interpretations. Also shown are the magnetostratigraphic correlations adapted from Chang et al. (2015). Black arrows show reconstructed paleoflow directions based on measurements of imbricated clasts.

Figure 3. Photos of typical carbonate lithologies in the study area. (A) Very fine grained marl, interpreted to be deposited in a palustrine environment, 350 m in section. (B) Paleosol with pedogenic carbonate nodules, 375 m in section. Note the uneven upper boundary, denoting subaerial erosions. (C) Shallow ground water carbonate nodules, 1680 m in section. Note the elongate bed-parallel shape, denoting influence of ground water.

Figure 4. Different types of primary carbonates in terrestrial environments: lacustrine limestone/marl, pedogenic carbonate, and shallow ground water carbonate cement (modified from Wright, 2007).

Figure 5. Photomicrograph (under cross-polarized light) of selected carbonate thin sections. Numbers in meters indicate their stratigraphic heights in the section. Carbonate type: C, carbonate cement; P, pedogenic carbonate; M, marl; L, limestone.

Figure 6. Stable isotopic variations ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) of carbonates from the Huatugou section, along with their stratigraphic heights. Four phases are recognized according to the oxygen isotopic variations. Green thick lines show trends of the most negative $\delta^{18}\text{O}$ values of carbonate cement; black solid curve and black dotted curve show trends of the most positive and most negative $\delta^{18}\text{O}$ values of marl/limestone, respectively.

Figure 7. Cartoons showing phases of topographic growth in the northern Tibetan Plateau, and how they controlled the oxygen isotopic variations in carbonates. See text for

869 detailed explanation.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7