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Termination of fluvial-alluvial sedimentation in the Xining Basin, NE Tibetan Plateau, and its subsequent geomorphic evolution

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A B S T R A C T

The Xining Basin in northern China lies topographically in the transitional zone between the Tibetan Plateau and the Loess Plateau, and climatically in the transitional zone between humid East Asia and the arid Asian interior. During the Cenozoic, the basin accumulated a thick, continuous sequence of fluviolacustrine-dominated red beds and subsequent eolian loess. The termination of sediment deposition in the Xining Basin is the key to understanding its subsequent evolution, the development of the modern Yellow River, and the history of dust deposition. Here we present the results of a detailed paleomagnetic study of the upper sedimentary sequence of the Xining Basin below the river terraces, and in addition we review previous studies of river evolution development. Our results indicate that the age of the uppermost part of the continuous sequence of fluviolacustrine-dominated red beds in the Xining Basin ranges from ca. 20 Ma to 4.8 Ma, suggesting that basin sedimentation ended at the earliest at 4.8 Ma, and they also provide an earliest age constraint for the subsequent river terrace development and eolian dust deposition. In addition, these results demonstrate that the termination of sedimentation in the Xining Basin and the nearby Linxia Basin was approximately coeval, suggesting the rapid regional tectonic uplift of the NE Tibetan Plateau and resulting reorganization of the landforms and development of the paleo-Yellow River in the late Pliocene.

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1. Introduction

Although the development of river terraces plays a critical role in revealing regional tectonic uplift and climatic change, they are often difficult to date. The NE Tibetan Plateau is highly suited for investigating the relationship between river terrace development and tectonic uplift and climate change because of the enormous topographic contrasts, strong tectonic uplift events, the presence of up to 16 river terraces, and thick (up to over 300 m) easily datable loess deposits covering the terraces. Moreover, the initiation of Huang He (Yellow River) development and subsequent major fluvial incision has exposed complete sequences of thick Cenozoic basin sediments (Figs. 1 and 2). As a response to later tectonic uplift, the intramontane basins of the region have undergone a rapid switch from continental sedimentation to fluvial incision and river terrace formation (Stokes, 2008; Ren et al., 2014; Zhang et al., 2017). However, either a relatively slow or a high uplift rate can inhibit terrace preservation (Veldkamp and Van Dijke, 2000) and interpreting the terrace records may be difficult in this region where the signature of climatic variations is superimposed on the spatial and temporal patterns of tectonically-driven rock uplift (Li, 1995; Pan et al., 2009; Zhang et al., 2014, 2017). Tectonic and climatic records from the sediments above or below the fluvial terraces of the region can provide important information on their generation, abandonment and subsequent preservation (Zhang et al., 2017). The greatest challenge for addressing these issues arises where there is no reliable chronology for the fluviolacustrine and eolian sequences. Dating of the loess sequences above the terraces takes advantage of this setting and provides an effective approach for constraining the ages of both the loess and the underlying terraces, and provides a base for further analysis of the tectonic uplift and climatic aridification (e.g., Burbank and Li, 1985; Li, 1995; J.J. Li et al., 1996; Z. Li et al., 1996; Li et al., 1997, 2014; Horton et al., 2004; Lu et al., 2004, 2007; Pan et al., 2009; Vandenberghe, 2008; Vandenberghe et al., 2011). This approach has been successful in dating the continuous Quaternary loess sequences on the terraces. However, there may be large uncertainties in dating older, discontinuous or eroded eolian deposits (Lu et al., 2004; Vandenberghe, 2008; Craddock et al., 2010; Vandenberghe et al., 2011).

The Xining Basin is an important area for conducting this type of study because it has accumulated a thick, continuous sequence of...
Fig. 1. (A) Geological map of the Xining Basin showing the distribution of the Cenozoic stratigraphy and locations of the studied sections (modified from Dai et al., 2006). GJS: Guanjiashan section; HW: Houwan section; CJB: Caijiabao section; XJ: Xiejia section; MJZ: Mojiazhuang section; blue star: Tashan borehole (TSB). The sketch map in the top right shows the location of the Xining Basin with respect to the Tibetan Plateau and Loess Plateau. (B) Digital elevation map (DEM) showing the geomorphology of the Xining Basin and the surrounding region. The gray-shaded areas indicate the extent of the modern catchments of the Xining Basin (I) and Qinghai Lake (II).
fluvial lacustrine-dominated sediments (Dai et al., 2006; Dupont-Nivet et al., 2007; Xiao et al., 2012) and up to 16 river terraces above the basin sediments have been cut by the Huang Shui River, a tributary of the Yellow River (Zeng et al., 1995; J.J. Li et al., 1996; Z. Li et al., 1996; Lu et al., 2004, 2007), and subsequent continuous eolian loess or red clay deposited on the terrace. Radiocarbon dating of organic matter,
tertholuminescence (TL) or optically stimulated luminescence (OSL) dating of mineral grains and paleomagnetic dating have been widely used to provide a chronological framework for the Quaternary loess sequences on the lower terraces of the Huang Shui River (Zeng et al., 1995, 1997; J.J. Li et al., 1996; Z. Li et al., 1996; Lu et al., 2004, 2007). However, it is difficult to constrain the ages of the higher terraces because the eolian dust deposits on these terraces are thin and discontinuous owing to erosion. This is likely to be the reason why there are large differences in dating results obtained by different authors, with ages ranging from middle Miocene (~16 Ma) to late Pliocene (~3.4 Ma) (J.J. Li et al., 1996; Z. Li et al., 1996; Lu et al., 2004; Vandenberghe et al., 2011; Wang et al., 2012, 2013, 2015). Nevertheless, the basin sediments directly beneath the river terraces, as well as the uppermost part of the basin sedimentary sequence, can provide maximum age constraints for the development of the higher terraces because river terraces can only develop after the end of basin deposition and the initiation of fluvial incision.

The Cenozoic stratigraphy in the Xining Basin is well characterized and is briefly summarized below. It comprises the Chinese standard mammal fossil unit of the Early Miocene Xiejian Stage (equivalent to the Aquitanian (MN 1–3) Stage of the European marine fossil zones) (Li et al., 1981; Qi et al., 1981; Qi and Qi, 1995; Deng et al., 2006). The deposits consist of unusually continuous fine-grained continental sediments of early Eocene to Pliocene age based on paleomagnetic dating and biostratigraphic studies (Dai et al., 2006; Xiao et al., 2012), which are well suited for reconstructing environmental and climatic changes (QBGM, 1991; Dai et al., 2006; Fang et al., 2015). Detailed paleomagnetic dating of the Xiejia section and drilling of its topmost part indicate an age range of 52.5–9.8 Ma for the stratigraphic sequence (Dai et al., 2006; Zan et al., 2015). The older part of this age range has been partially confirmed by paleomagnetic dating of the Shuiwan section in the central Xining Basin (Dupont-Nivet et al., 2007). However, no dates are available for the uppermost sequence of the Xining stratigraphy, and accurate dating of the sequence is essential for reconstructing the record of tectonic uplift and climate change. Recently, we found that this uppermost stratigraphic sequence occurs in the basin center around a narrow zone from Guanjiaashan-Caijabao to Mojiangzhang on which the highest terraces are developed (Figs. 1 and 2). Most recently, an important late Miocene Hipparion fauna was found in this uppermost sequence (Han et al., 2015; Yang et al., 2017) (Figs. 1 and 2), which provides precise dating and enables an robust age constraint for the highest terraces developed above. Correlation of marker beds provided by sedimentary structures and lithofacies from the previously studied magnetostratigraphic sections enables the ages of the sediments and the terraces to be estimated. However, to provide a reliable and high resolution chronology for the uppermost stratigraphic sequence and to redefine the age of onset of the river terraces and the geomorphological evolution of the Xining Basin, we conducted a detailed paleomagnetic study of the sedimentary sequence around Guanjiaashan-Houwan-Caijabao. In addition, we combined the results with our work at Mojiangzhang to provide a detailed chronology for the uppermost part of the Xining stratigraphic sequence, and to constrain the age of the oldest terrace development. Subsequently, we use the results to review and clarify the history of fluvial landform development and eolian dust depositional history in the region and to try to resolve the existing discrepancies.

2. Geological setting

The Xining Basin is situated in the transition zone between the Tibetan Plateau and the Chinese Loess Plateau, and has an elevation range of about 2200–3000 m. The basin is surrounded by the Datong Shan (Mts.), Laji Shan (Mts.) and the Riyue Shan (Mts.) reaching the elevations of about 3500–4800 m to the north, south and west (Fig. 1B), respectively, with the east open to the adjacent Minhe-Lanzhou Basins (Fig. 1). Two major WNW–ESE-striking faults, the North Central Qilian Shan Fault (NCQF) and the Laji Shan Fault (LSF), border the northern and southern margins of the Xining Basin (Fig. 1). The Huang Shui River, which flows roughly W–E through the basin center, has deeply incised the Precambrian to Mesozoic bedrock of the basin and formed several gorges and up to 16 river terraces in the basin. These terraces are mostly covered by eolian sediments (loess or red clay) with thicknesses ranging from tens of meters to over 200 m (Zeng et al., 1995; J.J. Li et al., 1996; Z. Li et al., 1996; Lu et al., 2007; Vandenberghe et al., 2011) (Fig. 2).

The Xining Basin is a flexural foreland basin or intracontinental foreland basin formed during the Paleogene between 55 and 45 Ma, probably in response to the India-Asia collision (Fang et al., 2007b; Dupont-Nivet et al., 2010). Since the Miocene, the tectonic deformation of the basin has been mainly controlled by two large faults (Fig. 1) (Dai et al., 2006; Vandenberghe et al., 2011; Yuan et al., 2013). Accelerated uplift, especially after 8 Ma, occurred simultaneously with continued regional deformation and brought about the gradual collapse of the flexure basin in the NE Tibetan Plateau, terminating the fluvial-alluvial sedimentation at about 3.6 Ma (Li, 1991; Fang et al., 2016). Three stronger uplift episodes called the Qingzang (3.6 Ma), Kunhuan (1.2 Ma), and Gonghe (0.15 Ma) movements occurred in the NE Tibetan Plateau (Li, 1995; J.J. Li et al., 1996; Z. Li et al., 1996; Li and Fang, 1999). Subsequently, the beveled (erosional) surfaces of a peneplain or pediplain was formed and started to be strongly incised and gradually enhanced by fluvial activity (Li and Fang, 1999; Pan et al., 2009; Vandenberghe et al., 2011).

The Cenozoic successions over a large part of the Xining Basin are gently deformed by NNE-trending thrust faults and folds, while the northern basin is dominated by E–W-trending structures (Dai et al., 2006). Paleontological and lithostratigraphic studies enable the Cenozoic stratigraphy in the Xining Basin to be divided into the Paleogene Xining and Neogene Guide Groups (Huang, 1977; Qi et al., 1981; QBGM, 1991). The Paleogene Xining Group is visually striking with thick gypsum beds and red-colored (brownish red to reddish brown) interbedded mudstones and siltstones. It is further divided into the Qijiachuan Fm., Honggou Fm. and Mahalagou Fm. based on the relative proportion of the gypsum beds compared to the reddish mudstones and siltstones, as first suggested by the Qinghai Oil Exploration Team (QOT) (Huang, 1977) and later accepted by the stratigraphic division of the Qinghai Bureau of Geology and Mineral Resources (QBGM) (QBGM, 1991; Dai et al., 2006). The Neogene Guide Group is generally characterized by the near absence of gypsum beds and by lighter colored (light reddish-yellowish brown-to-brownish-yellow) mudstones and siltstones with a coarsening-upward trend with an increasing proportion of interbedded grayish sandstone beds and finally thin conglomerate beds. It was further subdivided by the QOT and QBGM into four formations: the Xiejia Fm. (early Miocene), Chetougou Fm. (mid-Miocene), Xianshui Fm. (middle to late Miocene) and Linxia Fm. (Pliocene) (Huang, 1977; QBGM, 1991). In order to distinguish the same names of the stratigraphic divisions in the Lanzhou Basin (Zhai and Cai, 1984; Yue et al., 2001) and the Linxia basin (Xie, 1991), we renamed the Xianshui Fm. and Linxia Fm. as the Guanjiaashan Fm. and Mojiangzhang Fm. Detailed paleomagnetic dating of the Xiejia Section and the Tashan drilling core in the southern Xining Basin determined the age of the top of the Neogene Group to be 52.5–9.8 Ma, with ages of 52.5–30 Ma for the Xining Group and 30–9.8 Ma for the Guide Group (Dai et al., 2006; Zan et al., 2015). The ages of the included formations are: Qijiachuan Fm. – 52.5–50 Ma, Honggou Fm. 50–41.5 Ma, Mahalagou Fm. 41.5–30 Ma, Xiejia Fm. 30–23 Ma, Chetougou Fm. 22.5–18.2 Ma, and Guanjiaashan Fm. 18.2–9.8 Ma (Dai et al., 2006; Zan et al., 2015). For stratigraphic sequences younger than this section, we recently conducted paleomagnetic dating of the Mojiangzhang section (Yang et al., 2017) and subsequently other sections nearby (Guanjiaashan, Houwan and Caijabao sections) where the highest terraces of the Huang Shui River occur (T13–T15) (Fig. 2).

The highest terraces T13–T15 can clearly be seen and traced northwards with progressively increasing heights as follows: Guanjiaashan (T12), Houwan (T13), and Caijabao (T14) to Yanshang (T15) (Table 1)
and Fig. 2). These terraces are visually striking because of their partially carbonate-cemented gravel beds with an obvious imbricated structure and moderately sorted subrounded gravels consisting mainly of metasandstones with occasional quartzites and granites. Above these terraces are thick eolian deposits, the red clay and loess. The elevations of the top of the gravel beds of terraces T12 to T15 are 2680 m, 2740 m and 2766 m, respectively; with corresponding gravel bed thicknesses of 6 m, 3 m and 2 m (Table 1 and Figs. 2C and 3). Eolian red

<table>
<thead>
<tr>
<th>Terrace</th>
<th>Local name</th>
<th>Elevation of terrace gravel top (m)</th>
<th>Eolian loess-red clay thickness (m)</th>
<th>Age (Ma)</th>
<th>Dating method</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>T1</td>
<td></td>
<td>2241</td>
<td></td>
<td>0.01</td>
<td>$^{14}$C</td>
<td>J.J. Li et al., 1996; Lu et al., 2004; Zeng et al., 1995</td>
</tr>
<tr>
<td>T2</td>
<td></td>
<td>2250</td>
<td>8</td>
<td>0.05</td>
<td>$^{14}$C, TL</td>
<td>J.J. Li et al., 1996; Zeng et al., 1995</td>
</tr>
<tr>
<td>T3</td>
<td>Tuxiangdao</td>
<td>2296</td>
<td>30</td>
<td>0.12</td>
<td>TL</td>
<td>Lu et al., 2004</td>
</tr>
<tr>
<td>T4</td>
<td></td>
<td>2336</td>
<td>30</td>
<td>0.15</td>
<td>$^{14}$C, TL</td>
<td>J.J. Li et al., 1996; Zeng et al., 1995</td>
</tr>
<tr>
<td>T5</td>
<td>Beishansi</td>
<td>2380</td>
<td>42</td>
<td>0.54</td>
<td>Paleomagnetism</td>
<td>Zeng et al., 1995</td>
</tr>
<tr>
<td>T6</td>
<td></td>
<td>2400</td>
<td>90</td>
<td>0.67</td>
<td>Inferred from sedimentation rate</td>
<td>This study</td>
</tr>
<tr>
<td>T7</td>
<td></td>
<td>2436</td>
<td>120</td>
<td>0.78</td>
<td>Paleomagnetism</td>
<td>Zeng et al., 1995</td>
</tr>
<tr>
<td>T8</td>
<td></td>
<td>2490</td>
<td>160</td>
<td>1.17</td>
<td>Inferred from sedimentation rate</td>
<td>This study</td>
</tr>
<tr>
<td>T9</td>
<td>Dadunling</td>
<td>2508</td>
<td>232</td>
<td>1.2</td>
<td>Paleomagnetism</td>
<td>Zeng et al., 1997; Lu et al., 2007; H.Y. Lu et al., 2012</td>
</tr>
<tr>
<td>T10</td>
<td>Panzishan</td>
<td>2541</td>
<td>189</td>
<td>1.6</td>
<td>Paleomagnetism</td>
<td>Lu et al., 2007; H.Y. Lu et al., 2012</td>
</tr>
<tr>
<td>T11</td>
<td></td>
<td>2610</td>
<td>110</td>
<td>1.87</td>
<td>Inferred from sedimentation rate</td>
<td>This study</td>
</tr>
<tr>
<td>T12</td>
<td>Guanjiashan</td>
<td>2680</td>
<td>30</td>
<td>2.6</td>
<td>Paleomagnetism</td>
<td>This study</td>
</tr>
<tr>
<td>T13</td>
<td>Houwan</td>
<td>2708</td>
<td>106 (65 m L + 41 m R)*</td>
<td>3.3</td>
<td>Paleomagnetism</td>
<td>This study</td>
</tr>
<tr>
<td>T14</td>
<td>Caijiabao</td>
<td>2740</td>
<td>60 (20 m L + 40 m R)</td>
<td>3.6</td>
<td>Paleomagnetism</td>
<td>This study</td>
</tr>
<tr>
<td>T15</td>
<td>Yanshang</td>
<td>2766</td>
<td>38 (10 m L + 26 m R)</td>
<td>4?</td>
<td>Inferred from sedimentation rate</td>
<td>This study</td>
</tr>
</tbody>
</table>

L: loess; R: red clay; $^{14}$C dating of organic matter; TL: thermoluminescence.

Lithology and representative field photos of the Guanjiashan, Houwan and Caijiabao sections.

Fig. 3. Lithology and representative field photos of the Guanjiashan, Houwan and Caijiabao sections.
clay and loess directly overlie the river gravel beds of terraces T13–T15, but there is a remarkable undulating erosion surface between the red clay and the overlying loess. However, terraces T12 to T1 are only blanketed by loess, with the thickest loess deposits on T9 and T10, while the thickest loess layers on some of the lower terraces have been eroded and therefore are thinner (Table 1 and Figs. 2C and 3).

Much effort has been made to date the loess and red clay sequences using carbon-14 dating of organic material, TL/OSL dating of mineral grains, paleomagnetic dating, and correlation of paleosol-loess and magnetic susceptibility records with dated terrace loess sequences (Zeng et al., 1995, 1997; J.J. Li et al., 1996; Z. Li et al., 1996; Lu et al., 2004, 2007; H.Y. Lu et al., 2012; H.J. Lu et al., 2012). The resulting ages are listed in Table 1 and Fig. 2C.

Below these highest terraces are lower terraces, with a large height difference between each two adjacent terraces. Eolian red clay is absent from the lower terraces; instead, some of them are overlain by thick, yellowish, loose loess deposits (Fig. 2). Several drillings have been carried out on some of the thick loess sequences for paleomagnetic dating. The remainder of the lower terraces have thinner loess covers with generally well-exposed sequences that can be dated using carbon-14 dating of organic material and TL/OSL dating of mineral grains. The ages of the loess-covered terraces range from 1.6 Ma to the Holocene (Zeng et al., 1995, 1997; J.J. Li et al., 1996; Z. Li et al., 1996; Lu et al., 2004, 2007; H.Y. Lu et al., 2012; H.J. Lu et al., 2012). The resulting ages are listed in Table 1 and Fig. 2C.

3. Stratigraphy of the sampled new sections

The studied Guanjiashan, Houwan and Caijiabao sections are located in the central Xining Basin (Fig. 1B), on which the highest river terraces (T12–T13–T14) occur. The sections consist of two parts: the old fluviolacustrine red beds and the young terrace gravel deposits with overlying eolian (red clay-loess) deposits (Figs. 2 and 3). The stratigraphy of the basin deposits in the three sections is horizontal and the lithology is very similar, comprising the uppermost sequences of the Cenozoic basin strata. The succession is characterized by a coarsening-upward sequence of roughly cyclic, thick yellowish-brown mudstones and siltstones, with brown luvisol paleosols and sandstones with occasional fine conglomerates in the upper part of the basin fill sequence (Fig. 3). From the lithology and sedimentary structures of the three sections we identified three major sedimentary facies: braided river–floodplain, braided river with floodplain, and alluvial fan (Fig. 3). The terraces were formed after floodplain abandonment, and subsequently the dry terrace surfaces began to accumulate eolian dust (Pan et al., 2009). The terraces exhibit similar lithological sequences: Cenozoic fluviolacustrine sediments, clastic red beds in the depressions, and Paleozioc metamorphic rocks in the gorges that are covered by fluvial gravels of various thicknesses, intermingled with lenses of sands, silts, and clays and finally covered with eolian loess at the top (Vandenbergh et al., 2011), and the terrace gravels primarily consist of sandstone and quartzite, and are mainly rounded to sub-rounded and strongly weathered.

The Guanjiashan section (36°40.55′N, 101°50.41′E) is 336 m thick and consists of a coarsening-upward sequence from floodplain mudstones and sandstones to braided river sandstones and fine conglomerates, similar to the Guanjiashan, Houwan and Caijiabao sections, and terminates with thick alluvial fan pebble-cobble conglomerate beds (Fig. 3). Floodplain deposits dominate the lower section and consist of light brownish mudstones and siltstones with numerous paleosols or paleosol complexes and occasional thin shallow lake marl or thin river channel sandstone beds. The paleosols are pedogenically very similar to those of the Guanjiashan–Houwan–Caijiabao sections. In the middle parts of the section, an increased occurrence of sandstone beds intercalated with fine conglomerate beds exists, representing the encroachment of a braided river with strong flow force (cf. Fig. 3). The upper part of the section is dominated by thick pebble-cobble conglomerate beds interbedded with occasional sandstones and siltstones without paleosol development, which we interpret as coarser sediment input from an alluvial fan with stronger hydrodynamic force (cf. Fig. 3). All the measured conglomerate beds indicate a dominant southerly flow direction and similar gravel compositions to those of the Guanjiashan–Houwan–Caijiabao sections (cf. Fig. 3).

4. Sampling and measurements

The three studied sections include three adjacent terraces, T12, T13 and T14 (Figs. 2 and 3), and it is possible to compare their lithofacies and sedimentary facies to assess the spatial consistency of the chronology. Our overall aim is to date the sections in detail and to determine the onset of eolian dust deposition. Paleomagnetic samples were taken at 1–2 m intervals using a portable gasoline-powered drill equipped...
with a 2.5 cm diameter drill bit and a water-flushing system. All samples were oriented in situ using a magnetic compass. A total of 335 samples were obtained, 252 from the fluviolacustrine sediments below the terraces and 83 from the eolian deposits overlaying the terraces in the three sections. In addition, for cross-calibration purposes, each sample was cut into three parallel subsamples in the laboratory.

Stepwise thermal demagnetization was performed on two sets of specimens. Eighteen steps at intervals of about 20–50 °C up to 690 °C were used, with 8 to 12 steps at 20–50 °C intervals between 400 °C and 690 °C. The first set of specimens was measured on a 2G Enterprises Model 755 cryogenic magnetometer at the Institute of Tibetan Plateau Research, Chinese Academy of Sciences, in a magnetically shielded room (<150 nT). The second set of specimens was measured on a 2G Enterprises He-free u-channel SQUID magnetometer (2G-755R) at the Department of Geosciences, Tübingen University, Germany. The characteristic remanent magnetization (ChRM) was calculated by principal component analysis, using at least three consecutive steps trending towards the origin above 300 °C. Samples with maximum angular deviation (MAD) > 15° were not used for magnetostratigraphic interpretation. The characteristic remanent magnetization (ChRM) component was successfully isolated between 250 °C and 670 °C. After removal of a viscous magnetization below 200–250 °C the demagnetization paths decayed nearly linearly to the origin without a noticeable difference between the remanences carried by magnetite and hematite. Of the 335 samples measured, 277 samples gave reliable ChRM directions.

5. Results and discussion

5.1. Magnetic properties and mineralogy

Representative thermal demagnetization diagrams revealed two clear, rapid decreases in remanence intensity at around 580 °C and 650–670 °C (Fig. 4), indicating that magnetite and hematite are the main carriers of the remanent magnetization. The presence of these magnetic mineral phases was further confirmed by the κ-T curves which exhibit a sharp decrease in susceptibility near 580 °C and a further decrease at around 650–670 °C (Fig. 4), which are clear signatures of hematite and magnetite, respectively. It should be noted that the presence of hematite is usually masked by the much stronger magnetic contribution of magnetite (Dunlop and Özdemir, 1997).

Thermal demagnetization revealed that the Characteristic Remanent Magnetization (ChRM) component was successfully isolated between 250 °C and 670 °C. After removal of a viscous magnetization between 200–250 °C the demagnetization paths decayed nearly linearly to the origin without a noticeable difference between the remanences carried by magnetite and hematite. Of the 335 samples measured, 277 samples gave reliable ChRM directions.

5.2. Magnetostratigraphy

The accepted tilt-corrected ChRMs were analyzed separately for the eolian deposits and the fluviolacustrine sediments from the three sections. The mean directions of normal and reversed polarities are antipodal, indicating a primary record of the Earth’s magnetic field for all four groups (Fig. 5). A statistical bootstrap technique was used to test whether the ChRM vectors are Gaussian-distributed, and to assess the associated uncertainties for both normal and reversed ChRMs (Tauxe, 1998) (Fig. 5). For this test, the reversed polarities were flipped to their antipodes. For all four subsets, the 95% confidence limits of the two data sets overlap in all three components (Fig. 5), signifying that the reversed and normal modes cannot be distinguished at the 95% level of confidence, and indicating that all three sections pass the bootstrap reversal test (McFadden and McElhinny, 1990; Tauxe, 1998). Assessment of the remanence origin by a fold test is not possible for our data because the stratigraphy of all three sections is horizontal (Tauxe and Watson, 1994).

A jackknife technique (Tauxe and Gallet, 1991) was used to quantify the reliability of the magnetostratigraphy of the section. The jackknife parameters (J) calculated for the accepted specimen-mean directions have...
values of $-0.28$ in the Guanjiashan section, $-0.35$ in the Houwan section, and $-0.26$ in the Caijiabao section (Fig. 6), which all fall within the range of 0 to $-0.5$ recommended by Tauxe and Gallet (1991) for a robust magnetostratigraphic data set. This indicates that the sampling of the three sections has recovered $>95\%$ of the known polarity intervals.

The normal and reverse polarity intervals are designated $N1-N9/R1-R10$ in the Guanjiashan section, $N1-N9/R1-R9$ in the Caijiabao section, and $N1-N10/R1-R9$ in the Houwan section (Fig. 7). Correlation of the observed polarity sequences to the GPTS 2012 of Gradstein et al. (2012) is based on the following arguments (Fig. 7). The fossil mammals found in our section, together with the paleomagnetic dating of the Chetougou Fm. and the Guanjiashan Fm. in other sections nearby (such as the Xiejia-Tashan drilling core and the Mojiazhuang section) which also contain abundant fossil mammalian assemblages (Liu, 1992; Dai et al., 2006; Zan et al., 2015; Yang et al., 2017), provide robust...
of the Guanjiashan Fm. begins with a marked thick, followed by several striking brown paleosol complexes of luvic grayish-greenish carbonate-cemented sandstones and sandy marl, researchers (Li et al., 1981; Qiu et al., 1981)( Figs. 2 and 7). Near Cricetodon Miocene fossils can be directly traced to the nearby Houwan and Caijiabao sections. The values of J = 0.35 and −0.26 indicate that the section has recovered over 95% of the true number of polarity intervals.

constraints for the magnetostratigraphic correlation (Fig. 7). Two stratigraphic levels of fossil mammals were found at the bottom of the Guanjiashan Fm. by a local geological team. They are Artiodactyla indet and Rhinocerotidae indet, which are components of the middle Miocene fauna in the region (Liu, 1992) (Figs. 2, 3 and 7). The bottom of the Guanjiashan Fm. begins with a marked thick, fine-grained conglomerate or sandstone bed, which has clear parallel- or cross-beddings. In this marker bed in the other sections nearby, fossil mammals (Elasmotheriini, Comphotherium wimani and G. connexus) of middle Miocene age were found at Leidazhuang by the same local geological team (Liu, 1992) and at Diaogou near the Xiejia section by other researchers (Li et al., 1981; Qiu et al., 1981) (Figs. 2 and 7). In the Xiejia section, there are two horizons of fossil mammals including the well-known early Miocene Xiejia fauna in the Xiejia Fm. and the middle Miocene Cricetodon sp. in the basal sandstone bed of the Chetougou Fm. (Li et al., 1981; Dai et al., 2006) (Figs. 2 and 7). Near the top of the Guanjiashan section, there are other marker beds of grayish-green carbonate-cemented sandstones and sandy marl, followed by several striking brown paleosol complexes ofuvic cambisols above (Figs. 2, 3 and 7). These two bounding marker beds can be directly traced to the nearby Houwan and Caijiabao sections and are also well correlated to the Mojiazhuang section (Figs. 2, 3 and 7). In the uppermost parts of the Mojiazhuang section, we found late Miocene fossil mammals of Parelasmothierium sp., Hipparion dongxiangense and Chilotherium wimani (Han et al., 2015; Yang et al., 2017). Detailed paleomagnetic dating of the Mojiazhuang section has confirmed these biostratigraphic ages and determined the age of the section to be 12.7–4.8 Ma (Yang et al., 2017) (Fig. 7g).

With these robust constraints, all the observed polarity zones of the Guanjiashan, Houwan and Caijiabao sections are easily correlated with chron 2An-6n of the GPTS of Gradstein et al. (2012) (Fig. 7). The lower sandstone marker bed at the bottom of the Guanjiashan Fm. occurs within polarity interval N7 in the middle-lower part of the Guanjiashan section. The upper bounding marker beds of sandstones and paleosol complexes occur within the sequence from R2 to upper N1 in the upper part of the Guanjiashan section, and within R9-N9 and R9-N10 in the bottom parts of the Houwan and Caijiabao sections, respectively (Fig. 7). In the Guanjiashan section, we correlated the long, predominantly normal polarity intervals N1-N4 with chron 5An-5AnDn of the GPTS, N7 with chron 5Cn, and N9 with chron 5En-6n (Fig. 7b, c). The distinct long, predominantly normal polarity intervals N5-N8 in the Houwan section, and N6-N9 in the Caijiabao section, are correlated with the long predominantly normal polarity chron 4n-5n (Fig. 7d–f). Thus, our correlations constrain the ages of the fluvialocustrine red beds to 19.5–11.7 Ma (6n-5r) for the Guanjiashan section, 12.4–7.3 Ma (5An-3Bn) for the Houwan section, and 12.4–6.0 Ma (5An-3An) for the Caijiabao section (Fig. 7).

Because the thick continuous loess sequences on the lower terraces can be traced back to a minimum age of ~1.6 Ma on T10 (Zeng et al., 1995; Lu et al., 2004, 2007), and the age of the youngest fluvial-alluvial sequence was estimated to be about 4.8 Ma in the Mojiazhuang section (Yang et al., 2017) (Figs. 2 and 7g, h), the eolian deposits on the higher terraces T12 to T15 were formed between 4.8 and 1.6 Ma. These constraints imply the following correlations: R1 in the Guanjiashan section with chron 2r (Matuyama reversed polarity interval below the Olduvai event); R1, N1-N4 and R4 in the Houwan section with chron 2r and 2An.1n and 2An.1r (Kaena event); R1 and N1-N4 in the Caijiabao section with chron 2r and 2Cn (Gauss chron); and R2 with the Kaena to Mammoth events. These correlations indicate that the ages of T12, T13 and T14 are about 2.6 Ma, 3.2 Ma and 3.6 Ma, respectively (Fig. 7).

Based on the depth-versus-age distribution of polarity changes for the fluvialocustrine sediments of the five sections, there was a significant change in the sedimentation rate at about 8.0 Ma, with rates of about 4.5 cm/ka before, and 2.3 cm/ka after this datum (Fig. 8). This change in the sedimentation rate corresponds well to a significant increase in the conglomerate and sandstone content after about 8 Ma, which further supports our magnetostratigraphic interpretations.

5.3. Termination of sediment deposition in the Xining basin and subsequent major fluvial incision, and their constraints on eolian dust accumulation

The changes in basin sedimentation and fluvial incision can be attributed to a combination of several factors, such as rock uplift, changes in runoff caused by changes in climate and precipitation, and changes in the base level and the river gradient (Li et al., 1997; Miao et al., 2008; Zhang et al., 2014, 2017). The importance of the individual factors is likely to have varied within the Xining Basin since the Miocene and cannot be clearly separated in our study. However, it is noteworthy that the Huangshui River terraces are well-preserved on the northern side of the river but not on the southern side (Fig. 2B), suggesting the uplift of the eastern Qilian Shan to the north of the Xining Basin plays a major role; the narrow gorges cut by the river in the western basin and those cut by its tributaries in the northern basin are deeper than those in the eastern basin, indicating stronger basement uplift closer to the marginal mountains (Riyue Shan and eastern Qilian Shan) of the basin. Some small scale thrust-folds and dipper strata angles developed along the western and northern margins of the basin (Fig. 2B, C), indicating compression exerted by the uplift of the basin-marginal mountains. Furthermore, the variations of the incision rate in the Xining Basin and Linxia Basin (Li et al., 2014) (Fig. 9A, B) are in phase with those of the sedimentation rate and the conglomerate content from the surrounding area (Fang et al., 2005a; Li et al., 2014) (Fig. 9C), which are confirmed as representing abrupt tectonic events in different basins throughout the NE Tibetan Plateau (Bloemendal and DeMenocal, 1989; Yin et al., 2002; Fang et al., 2005a, 2005b, 2007a; Wang et al., 2006). However, these changes are not consistent with the persistent aridification trend evident in the pollen record (Ma et al., 2005; Wu et al., 2007, 2011; Li et al., 2014) (Fig. 9D), changes in sea-level (Haque et al., 1987) (Fig. 9E), and long-term global cooling since the early Pliocene (Zachos et al., 2008) (Fig. 9F). Based on fundamental principles, the combined factors of tectonism, erosion and weathering are responsible for the development of a basin system, including the surrounding mountains and fluvial and lake systems (Fig. 10A). Tectonic uplift with strong compression was responsible for driving headward erosion and reversal of the flow.
direction, lake shrinkage and even depocenter migration (Fig. 10B).

With continuing tectonic uplift, the river should have commenced strong incision given the fall in base level, eventually forming multiple terraces and accumulating loess (Fig. 10C). This model of fluvial incision and river terrace formation resulting from uplift is also confirmed in other basins external to the Tibetan Plateau (Leeder and Alexander, 1987; Lavé and Avouac, 2000; Stokes, 2008; Pan et al., 2012; Ren et al., 2014; Zhang et al., 2017). These observations suggest that pulsed tectonic uplift rather than climatic perturbations was the major driver of sedimentary changes and river terrace development in the study area. In response to subsequent tectonic uplift, the intramontane basins rapidly switched from continental sedimentation to fluvial incision and river terrace formation (Stokes, 2008; Ren et al., 2014; Zhang et al., 2017).

The coarsening-upward sequences in all the studied sections are associated with an increase in the sedimentation rate (Fig. 8) and with lithofacies changes from an early floodplain-dominated sedimentary
environment to a braided river system and finally to an alluvial fan environment (Figs. 3 and 7), indicating that the surrounding mountains began to be uplifted in the late Miocene. Accompanied by a large change in the sedimentation rate (Fig. 8), the uplift accelerated after about 8 Ma, which is confirmed by changes in sedimentary provenance, sediment coarsening and continuous clockwise block rotations throughout.
section, and suggest the fluviallacustrine sediments between 6.0 Ma and 3.6 Ma were probably eroded because of the strong tectonic uplift at about 3.6 Ma. Thus, the rapid uplift and deformation must have occurred between 3.6 Ma and 4.8 Ma. This is almost the same age as that obtained for the strong tectonic event called ‘phase A’ of the Qingzang Movement, dated to 3.6 Ma, in the nearby Linxia Basin to the east (J.J. Li et al., 1996; Z. Li et al., 1996; Fang et al., 2003) and the Guide Basin just to the south (Fang et al., 2005a) (see Fig. 1 for locations). In summary, the termination of deposition and onset of deformation of the Xining Basin sediments, and the subsequent origin of the paleo-Huang Shui River, was earlier than the entire occurrence of the Huanghe River at about 1.2 Ma in the NE Tibetan Plateau (Pan et al., 2009; Hu et al., 2011; Zhang et al., 2014), and reflects the rapid uplift of the NE Tibetan Plateau.

The successive development of the lower terraces records the subsequent incision history of the Huang Shui (Fig. 2). The most significant event during this later history of the Huang Shui River is the reversal of the flow direction from westward to eastward at about 1.2 Ma at terrace T9 (Dadunling) (Miao et al., 2008). We interpret this flow reversal, which was probably caused by river capture and beheading (Bishop, 1995), as reflecting the rise of the Riyue Shan to the west of the basin at about 1.2 Ma. This means that prior to 1.2 Ma the paleo-Huang Shui River flowed into Qinghai Lake to the west of the Riyue Shan. Tectonic investigation of the Riyue Shan shows that the dextral Haiyan fault was probably active in the mid-Pleistocene and induced an extensional subsiding valley where the Yao Shui River (source of the modern Huang Shui River) flowed along the valley into the Xining Basin and subsequently incised to a depth of about 300 m (Wang et al., 2011).

Comparisons of the terrace elevation, terrace gravel relative height and incision rate of the Huang Shui in the Xining Basin and Daxia Hang-Huang He in the Linxia Basin demonstrates that intervals of rapid incision occurred at about 1.8 Ma, 1.2 Ma, 0.6 Ma and 0.13 Ma (Fig. 9A, B), suggesting regional stepwise rapid uplifts of the NE Tibetan Plateau at those times (Li et al., 2014).

This configuration at the end of deposition in the Xining Basin, and the subsequent incision and terrace formation, constrains the age of the eolian deposits (red clay) on the higher terraces to about 3.6–2.6 Ma (late Pliocene). Our predominantly normal-polarity paleomagnetic pattern of the red clays on terraces T13 (Houwan) and T14 (Caijiabao), correlated to the Gauss Normal Chron of the GPTS (Fig. 7), confirms this constraint. Thus, the change from the well-developed luvisol paleosols of cambisols that now occur only in East China, with numerous large mammals typical of forest-steppe environments in the late Miocene, to the fine eolian deposits, the red clay of the late Pliocene, and finally to the coarse Quaternary loess, suggests an intensification of the aridification of the Asian interior and associated weakening of the East Asian summer monsoon since the late Miocene.

6. Conclusions

We have investigated and characterized the uppermost parts of the Cenozoic sedimentary sequences in the Xining Basin, which extends the continuous stratigraphic record from the previous Eocene-middle Miocene to the late Miocene-early Pliocene. Our findings highlight the great potential of the region for reconstructing paleoclimatic changes and landform evolution. Detailed paleomagnetic dating of the uppermost successions yields an age range of about 20 Ma to 4.8 Ma and constrains the age of the highest terraces and their overlying eolian deposits (red clay) to about 3.6–2.6 Ma. Synthesis of the evolution of the Huang Shui River based on this and our previous studies reveals a stepwise incision history that indicates the occurrence of rapid pulses of NE Tibetan Plateau uplift at about 3.6 Ma, 2.6 Ma, 1.8 Ma, 1.5 Ma, 1.2 Ma, 0.8 Ma, 0.5 Ma, and 0.13 Ma. These ages are in good agreement with the previously documented record of episodic regional tectonic activity, such as the QingZang, Kunhuang and Gonghe Movements.
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