Magneto- and litho-stratigraphic records of the Oligocene-Early Miocene climatic changes from deep drilling in the Linxia Basin, Northeast Tibetan Plateau

Fuli Wu, Xiaomin Fang, Qingquan Meng, Yan Zhao, Fenjun Tang, Tao Zhang, Weilin Zhang, Jinbo Zan

a CAS Center for Excellence in Tibetan Plateau Earth Sciences, Key Laboratory of Continental Collision and Plateau Uplift, Institute of Tibetan Plateau Research, Chinese Academy of Sciences, Beijing 100101, China
b School of Earth Sciences, Key Laboratory of Western China's Mineral Resources of Gansu Province, Lanzhou University, Lanzhou 730000, China
c University of Chinese Academy of Sciences, Beijing 100049, China

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ABSTRACT

The East Asian monsoon is generally regarded to have initiated at the transition from the Late Oligocene to the Early Miocene. However, little is known about this process because of a lack of continuous strata across the boundary between the Late Oligocene and the Early Miocene in Asia. Based on previous drilling (core HZ-1) in the Miocene sediments in the southern Linxia Basin in NW China, we drilled a new 620 m core (HZ-2) into the Late Oligocene strata and obtained 206 m of continuous new core. The detailed paleomagnetism of the new core reveals eleven pairs of normal and reversed polarity zones that can be readily correlated with chrons 6Bn-9n of the geomagnetic polarity time scale (GPTS), define an age interval of 21.6–26.5 Ma and indicate continuity from the Late Oligocene to Early Miocene. The core is characterized by the remarkable occurrence of brownish-red paleosols of luvic cambisols (brown to luvic drab soils) above reddish-brown floodplain siltstones and mudstones, which suggest that the East Asian monsoon likely began by 26.5 Ma. In contrast to the siltstone and mudstone of the Late Oligocene strata, the Miocene strata begin with a thick fine sandstone bed, which marks sudden increases in erosion and loading that most likely reflect a response to tectonic uplift. The hematite content and redness index records of the core further demonstrate that the monsoonal climate in the Late Oligocene to Early Miocene in this area was mainly controlled by global temperature trends and events.

1. Introduction

The Oligocene was a period of climatic deterioration that witnessed the initiation of the continent-wide ice sheet on Antarctica and the formation of the Antarctic circumpolar current. Deep-sea δ¹⁸O records (e.g., Miller et al., 1987; Zachos et al., 2001) and terrestrial palynomorphs recovered from Antarctic margin drill holes (e.g., Mildenhall, 1989; Raine, 1998; Raine and Askin, 2001) indicate climatic amelioration in the Late Oligocene, which may be attributable to deep-sea warming coupled with a collapse of the Antarctic ice sheet (e.g., Miller et al., 1987; Zachos et al., 2001). Subsequently, until the middle Miocene, the global ice volume remained low, and the bottom water temperatures increased slightly but were punctuated by several short intervals of glaciations (called Mi-events) (Flower and Kennett, 1994; Zachos et al., 2001). Therefore, the period from the Late Oligocene to the Early Miocene was a critical interval for the climate transition and change in the Cenozoic era.

In Asia, evidence from both temporal and spatial investigations shows that a climate transformation from a zonal pattern to a monsoon-dominated pattern occurred roughly from the Late Oligocene to the Early Miocene (Liu et al., 1998; Guo et al., 2002, 2008; Wang et al., 2003a, 2003b, 2005; Sun and Wang, 2005; Wang, 2009; Qiao et al., 2006; Qiang et al., 2011), though several studies suggest that the East Asian monsoon likely began by 26.5 Ma. In contrast to the siltstone and mudstone of the Late Oligocene strata, the Miocene strata begin with a thick fine sandstone bed, which marks sudden increases in erosion and loading that most likely reflect a response to tectonic uplift. The hematite content and redness index records of the core further demonstrate that the monsoonal climate in the Late Oligocene to Early Miocene in this area was mainly controlled by global temperature trends and events.
of well-dated and continuous terrestrial sediment sequences that record the climate changes across the boundary between the Oligocene and Miocene.

The Linxia intracontinental foreland basin (Fang et al., 2016) provides an ideal opportunity for this purpose. It is located in a transitional zone between the East Asian monsoon and non-monsoon Asian inland dry climate (Fig. 1). Detailed stratigraphic, biostratigraphic and magnetostratigraphic studies suggest that the exceptionally long continuous sedimentary succession in this basin was deposited from the Oligocene to the Pliocene (Li et al., 1995; Fang et al., 2003, 2016; Deng et al.,
In particular, the Heilinling (HLD) section (Figs. 1, 2) in the foredeep of the southern Linxia basin includes continuous thick floodplain sediments with three levels of fossil mammals but few lithofacies variations, so it is suitable for revealing climate change (Fang et al., 2016). Unfortunately, the stratigraphy at Heilinling outcrops over a thickness of only 226.5 m. Therefore, in 2010 and 2011, we drilled a 129.2 m core (HHT) at the top of the section and a 477.7 m core (HZ-1) (Fig. 2) at its bottom. Detailed paleomagnetic measurements demonstrated an age of 23.3 Ma for the bottom of core HZ-1 (Fang et al., 2016). Encouraged by those results, we performed additional deeper drilling (core HZ-2, drilling depth 620 m) at the same site as core HZ-1 in December 2015 using the same drilling technology (three-in-one tube drilling, in which the core is directly recovered into an inner plastic tube for better protection and recovery rate) and obtained 206 m of new core with an recovery rate of over 95% (Fig. 2). In this paper, we carried out a detailed paleomagnetic analysis of this core and analyzed its lithofacies, hematite contents and redness proxies. Based on these data, we attempted to reconstruct the climatic changes across the boundary between the Late Oligocene and the Early Miocene.

2. Geological setting and stratigraphy

The Linxia Basin is located on the northeastern margin of the Tibetan Plateau and has an average elevation of 2000–2600 m above sea level (asl). It is surrounded by several high mountains, including the West Qin Ling Mts. to the south, which have a maximum elevation of 3767 m, the Laji Shan (Mts.) (4524 m asl) to the west, and the Maxian Shan (Mts.) (ca. 3000–3500 m asl) to the north, and it opens to the east to the large Longzhong Basin (Fig. 1). The modern climate in the Linxia Basin exhibits the features of the continental monsoon with a mean annual temperature (MAT) of 6.7 °C, an average monthly maximum temperature of 18 °C and mean annual precipitation (MAP) of > 500 mm. Warm temperate zone montane broadleaved forests and subalpine coniferous forests are distributed in the mountainous areas around the basin, and the landscape within the basin is temperate broadleaved deciduous forest-grassland.

The Cenozoic stratigraphy of the Linxia Basin mainly consists of red beds of fluvio-lacustrine mudstones and siltstones in the center and thick strata of gray to reddish-brown conglomerates and sandstones on the margins (Fang et al., 2003, 2016). Previous studies have divided the stratigraphy into eight formations based on lithofacies, contacts, palaeontology and magnetostratigraphy, including the Late Oligocene Talu (TL) Fm., the Early Miocene Zhongzhuang (ZZ) Fm., the Middle Miocene Shangzhuang (SZ) Fm., the Late Miocene Dongxiang (DX) Fm. and Liushu (LS) Fm., the Early and Late Pliocene Hewangjia (HWJ) Fm. and Jishi (JS) Fm. or Jishi Conglomerate Bed and the Early Pleistocene Dongshan (DS) Fm. (Li et al., 1995, 1997; Fang et al., 2003; Fang et al., 2016) (Fig. 2).

The HLD section (35.38°N, 103.32°E) is located on the southern margin of the Linxia Basin and consists of the HZT core (129.2 m) in the upper part (0–129.2 m), the HLD outcrop (226.5 m) and the HZT-1 core (477.7 m) in the middle and lower parts (120–824.5 m) (Fang et al., 2016) and the HZT-2 core (620 m) in the lowermost part (761–967.15 m) (Fig. 2). This paper studied the intervals of the section that include the TL Fm. and ZZ Fm. (Fig. 2). The TL Fm. and upper ZZ Fm. are composed of an upward-fining cycle of alternating brownish siltstones and reddish-brown mudstones (Fig. 2II-a) and strongly to moderately developed calcic paleosol complexes (Fig. 2II-b and e), which are mostly interpreted as overbank floodplain deposits (Fig. 2). These paleosols are luvis paleosols that are characterized by a luvis Bt horizon or a cambic Bw horizon, an underlying distinct calcic Bk horizon, and occasionally an additional calcic Ck horizon (Fang et al., 2016). The Bt horizon is approximately 0.5–1 m thick, more reddish in color, has black Fe/Mn mottling stains and contains a well-developed granular ped structure (Fig. 2II-b and e). The Bw horizon is similar to the Bt horizon but is lighter red (mostly dull reddish-brown) and has less pedality (ped to block structures). The Bk horizon is located immediately below the Bt or Bw horizon, is light brownish gray or yellowish-gray, and contains many white carbonate coatings, small nodules and imregnations that cement the sediments (Fig. 2II-b). Down-leaching carbonate occasionally reaches the soil’s parent materials (mudstone or siltstone) below the Bk horizon, but its content is less than that in the Bk horizon, and it forms a Ck horizon (Fig. 2II-I). The lower ZZ Fm. is composed of alternating thick reddish-brown parallel-bedding fine sandstones (Fig. 2II-d) and massive siltstones, which are classified as a river channel facies over a floodplain (Fig. 2) (Fang et al., 2016).

3. Sampling and methods

Core HZ-2 can be linked to core HZ-1 by the sandstone marker bed (Fig. 2I). We collected samples starting at a depth of 414.5 m in core HZ-2, which overlaps with core HZ-1 by 63.7 m. Paleomagnetic samples were collected at intervals of 0.5–1 m within the overlapping section and at 0.25–0.5 m intervals elsewhere, and a total of 554 block samples were collected (Fig. 2I). Each block sample was cut into two 2 × 2 × 2 cm specimens for measurements. One set of specimens was subjected to stepwise alternating-field (AF) demagnetization in 15 steps (0, 2, 4, 8, 12, 16, 20, 30, 40, 50, 60, 80, 100, 120 and 140 mT). For cross-checking, 80 samples from the second set were systematically thermally demagnetized in 19 discrete steps between room temperature and 690 °C (intervals of 50 °C below 550 °C and 10–30 °C above 550 °C) in a Magnetic Measurement Thermal Demagnetizer (MMDT80). All of the remaining measurements were then conducted using a 2G Enterprises Model 760-R cryogenic magnetometer installed in a field-free space (< 100 nT) at the Institute of Tibetan Plateau Research, Chinese Academy of Sciences.

180 samples were taken at 1–2 m intervals below a depth of 600 m (samples over 824 m were collected from core HZ-1, the else from core HZ-2) for diffuse reflectance spectroscopy (DRS) analysis for redness and hematite content measurements (Barranco et al., 1989; Balsam and Deaton, 1991; Deaton and Balsam, 1991). Most of the samples were collected from mudstones or siltstones so the impact of the sediment matrix could be eliminated from the results. First, all of the samples were ground in an agate mortar. The finely powdered samples were then pressed by hand into 20–40 mm rectangular plastic holders. The reflectance spectra of the samples were measured in a Purkinje General TU1901 UV-vis spectrophotometer with a diffuse reflectance attachment (reflectance sphere) from 360 to 850 nm at 1 nm intervals. The DRS analysis was conducted at the Institute of Tibetan Plateau Research, Chinese Academy of Sciences. All of the reflectance data were analyzed by the first derivative test to calculate the hematite contents (Balsam and Deaton, 1991; Deaton and Balsam, 1991).

4. Magnetostratigraphy of core HZ-2

The natural remanent magnetization (NRM) of the samples from core HZ-2 is within the range of 0.1–10 mA/m, and most of them are within the range of 1–10 mA/m (Fig. 3). The remanent magnetization (RM) of all of the samples decreases as the intensity of the applied magnetic field increases. When the intensity of the applied magnetic...
field reaches 140 mT, the RM of most of samples decreases to within 25% of the NRM (Fig. 3a–d), and the RM of some of the samples decreases to within 25%–50% of the NRM (Fig. 3e, i and k). There are two types of vector graphs for the RM direction. In the first type, there are two RM directions, and the RM direction shows remarkable changes at turning points of 8 mT or 12 mT (Fig. 3a and k) and 30 mT (Fig. 3b and c), which indicates the elimination of secondary RM. After the turning points, the RM direction approaches the origin and can represent the stable characteristic remanent magnetization (ChRM) directions. In the other type of vector graph, the RM direction does not change significantly (Fig. 3d, e and i), and the stable ChRM direction can be acquired. The higher ChRM values after 140 mT AF-demagnetization of...
the samples indicate that some of the ChRM is carried by hard magnetic minerals, such as hematite. The RM of the thermally demagnetized samples decreases significantly at 550–580 °C (Fig. 3h) and 610 °C (Fig. 3f), and the most significant RM decrease occurs at temperatures higher than 650 °C (Fig. 3f–h, j and i), all these indicate that the main RM carriers in the samples are magnetite and hematite. Due to the existence of hematite with a high coercive force, complete demagnetization cannot be achieved through alternating field demagnetization. The RM of a few samples increases before 120 °C (Fig. 3f) but then decreases with increasing temperature, which suggests the elimination of the viscous remanence carried by goethite. For thermal demagnetization, the RM direction generally approaches the origin at temperatures higher than 400 °C (Fig. 3f and g), and the RM direction approaches the origin at temperatures higher than 610 °C in some samples (Fig. 3h), which can represent the ChRM direction. The demagnetization effects of the samples are generally the same as those of the samples from core HZ-1 (Fang et al., 2016).

The directions of the ChRM components of all of the paleomagnetic samples from core HZ-2 are calculated through principle component analysis (Kirschvink, 1980), and the results are generally the same for the alternating field demagnetization and thermal demagnetization (Fig. 3i, j, k, and l), which substantiates the reliability of the test results. Because the core azimuth was not controlled during sampling, the magnetic declination has no significance, and the geomagnetic polar column can only be created with geomagnetic inclination data (Fig. 4). During sample selection, the samples with maximum angular deviations (MAD) > 15° that were used to determine the ChRM direction with fewer than four points were excluded. A total of 52 samples, which account for 9.4% of the samples, were excluded. Because the sampling was conducted at high resolution and the excluded samples were not concentrated in contiguous layers, the precision and interpretation of the paleomagnetic polar columns was not affected. Based on the principle that at least two samples of the same polarity define each polarity event, 54 samples (account for 9.7% of the samples) were excluded, then 11 positive polarity events (N1–N11) and 11 negative polarity events (R1–R11) were obtained (Fig. 4d).

Fig. 4 shows that the polarity zones (R1–R3, N1–N2) obtained from the interval of core HZ-2 that overlaps with core HZ-1 are highly consistent with those (R38–R40, N39–N40) of core HZ-1 (Fang et al., 2016), which confirms the connection between the cores. This provides a reference for determining the age of the upper boundary of core HZ-2 because the correlation of magnetic polarity zones of core HZ-1 with the GPTS is very robust, with considerable age constraints from three levels of fossil mammals that were found in the section (Fang et al., 2016). Therefore, we can readily correlate normal polarity zone N1 of core HZ-2 with chron 68n and reversed polarity zone R2 with chron 6Br of the GPTS (Hilgen et al., 2012) (Fig. 4f, g). Thus, we correlate N3 and N4 with chron 6Cn.2n and 6Cn.3n, respectively, considering that the corresponding channel sands indicate high erosion and sedimentation rates (Fig. 5) in comparison with the underlying floodplain mudstones and paleosols (Fig. 4). The frequent occurrence of long normal zones N6 to N10 can be correlated with chron 7Cn to 8n, and N11 is readily correlated with the top part of 9n (Fig. 4f, g). The deposition rate of core HZ-2 established by the correspondence between the depth and the magnetostatigraphic age (Fig. 5) shows it ranges stably between 2.64 and 6.33 cm/ka, and the average deposition rate is 4.28 cm/ka, which is similar with the one at bottom of the core HZ-1 (3.84 cm/ka) (Fang et al., 2016). Then this correlation demonstrates that the age interval of core HZ-2 is 21.6–26.5 Ma based on extra-polations with the sedimentation rate (Fig. 5). The results indicate continuity in the stratigraphic record from the Late Oligocene to the Early Miocene and shows that the beginning of the Miocene is marked by the appearance of the sandstone bed above the reddish mudstones and paleosols at approximately 23 Ma, which is consistent with the standard age for the start of the Miocene within the errors (Cohen et al., 2013).

5. Variations of lithology, hematite content and redness and their indications of Asian monsoon change

5.1. DRS measurements and variations of hematite content and redness records

DRS analysis is highly sensitive to iron oxide minerals in the soil and sediments and has been recognized as an important instrument to identify and estimate iron oxide minerals in soil and sediments (Schwertmann, 1971; Cornell and Schwertmann, 1996; Balsam et al., 2004; Fang et al., 2015; Torrent et al., 2006; Nie et al., 2017). Because the characteristic reflectance spectra are different for different iron oxide minerals, several iron oxide minerals in the samples, especially hematite and goethite, can be identified by DRS scanning. The DRS first order derivative of the cores HZ-1 and HZ-2 has only one significant peak, and the center is at 568 nm, which indicates the content of the hematite (Fig. 6, Fang et al., 2015). The redness is the percentage ratio of the reflectivity of the red band at 630–700 nm to the total reflectivity of the visible band (400–700 nm) (Judd and Wyszecki, 1975).

Fig. 7a shows the variations of hematite content and redness for the stratigraphic interval of 600–976 m (18–26.5 Ma) in the section composed of cores HZ-1 and HZ-2. They show a generally homogeneous pattern with high but narrow ranges of 0.26–0.45 for the hematite content and 31.55–42.59 for the redness, which are punctuated by several small but clear short-lasting valleys (lower values) at depths of 820 m (22.9 Ma), 773 m (21.9 Ma), 726 m (21 Ma), 669 m (19.7 Ma), 639 m (18.9 Ma) and 610 m (18.1 Ma), which are marked as hematite and redness events HR 1 to HR 6, respectively (Figs. 4b, c and 7a, b).

5.2. Changes in the Asian monsoon from the Late Oligocene to the Early Miocene

The studied stratigraphic intervals of cores HZ-1 and HZ-2 are dominated by fine sediments of reddish-brown mudstones and siltstones, which are intercalated with many well developed paleosols and a short interval of fine sandstones between them that typically characterize a floodplain environment (Reading, 2009) (Figs. 2 and 4a). We interpreted the well-developed luvic or cambic paleosols, which only form under subaerial conditions, and the subaerially oxidized floodplain mudstones and siltstones as an indication that the East Asian monsoon had formed by approximately 26.5 Ma and throughout the studied Late Oligocene to Early Miocene interval. This is because each kinds of soils represent a specific climatic condition (Retallack, 2001), and the modern soils such as these luvic or cambic paleosols form in easternmost China like Shandong Province (Fig. 1a), where the MAT and MAP are approximately 12–14 °C and 600–850 mm, respectively, the warmest month temperature is 23–27 °C, and the average precipitation in the summer (June to August) accounts for > 60% of the MAP (Xiong and Li, 1987). In other words, the climatic condition indicated by the paleosols is better than that at the modern Linxia Basin. Moreover, the water vapor source at that time cannot be the westerly wind, as many studies show that the Northwest of the China is more arid after the Oligocene (e.g. Dupont-Nivet et al., 2007; Xiao et al., 2010; Sun et al., 2014, Sun and Windley, 2015), the water vapor can only be brought in by the East Asian monsoon, and the monsoon in the Linxia Basin was stronger in the Late Oligocene and Early Miocene than at present.

The abundant mammal fossils that have been discovered in the Linxia Basin provide further support for this interpretation (Deng et al., 2004, 2013). The giant rhino fauna in the Linxia Basin in the Late Oligocene mainly include Daungariotherium orgasmos, Tsaganomys altaicus, Megalapterodon sp., Schizotherium ordosium, Tripolopus sp., Ardynia sp., Ardynia sp. nov., Allacrepes sp., Paraceratherium sp., Ronztherium sp., Aprotodon lanzhouensis, and Paracentelodon macrognathus, and this mammalian assemblage indicates a warm and humid habit (Deng et al., 2004, 2013), which confirms our interpretation of the monsoon.
Fig. 4. Depth functions of the lithology (a), changes of hematite content and redness (b and c), and inclination directions of core HZ-2 and the determined polarity zones (black/white bars indicate normal/reverse polarity) (d, f). For comparison, the magnetic inclination directions and the determined polarity zones of the lower part of core HZ-1 are also plotted (e, f) (Fang et al., 2016). Both polarity zones are correlated with the GPTS of Hilgen et al. (2012) (g).
The lower section of the ZZ Fm. consists of alternating parallel-bedded sandstones and siltstones that belong to a river channel facies (Fig. 2). In contrast to the fine siltstone and mudstone of the TL Fm., these sandstones appeared at the beginning of the Miocene and indicate sudden increases in erosion and loading. These were likely a response to the Early Miocene tectonic uplift of the NE Tibetan Plateau (Fang et al., 2003, 2016) because this event has been widely reported to have occurred on and around the Himalaya-Tibetan Plateau (Yin et al., 1998; Wang et al., 2003a, 2003b, 2015; Fang et al., 2005; Sobel et al., 2006; Lu et al., 2012; Lease et al., 2011; Wang et al., 2016).

We also consider the hematite content and redness to be sensitive climatic proxies. Previous studies show that hematite is a common mineral in soils and sediments and is a product of silicate minerals that have weathered into the soil (Schwertmann, 1971; Balsam and Damuth, 1991).

Fig. 5. Age versus stratigraphic depth plot of the core HZ-2 using the correlations of the magnetozones shown in Fig. 4f. GPTS is cited from Hilgen et al., 2012.

Fig. 6. First derivative curves for several samples and the highest first derivative values at 568 nm from cores HZ-1 and HZ-2. The dark line indicates the first derivative for 100% hematite (Balsam and Deaton, 1991).
In the period of the 26.5–18 Ma, the main provenance of sediments in the HLD section did not change (Fang et al., 2016). Even though we could not rule out the impact of original hematite in the rocks, especially Permian-Triassic igneous rocks in the West Qinling to the south of the Linxia Basin (Liu et al., 2015), we think it would not exert a significant role in our hematite record in million and less million scales because its concentration in the rocks of the source areas is very low, chemically resistant (White and Buss, 2014) and hard to impact short times of climatic changes. Rather, the hematite in the sediments of the basin should mostly be formed in the depositional stage with climate changes (e.g., Hu et al., 2005; Fang et al., 2015). Moreover, hematite is mainly distributed in soils with good oxidizing conditions in tropical and subtropical zones and therefore mainly indicates a relatively hot environment in the modern (Schwertmann, 1971; Comell and Schwertmann, 1996; Fang et al., 2015; Ji et al., 2002, 2006, 2007). The relationship between color and climate is complicated, but most studies indicate that color is related to the contents of organic matter and coloring minerals. For example, redness is usually closely related to hematite content (Fang et al., 1999, 2015; Giosan et al., 2002, White et al., 2007). Therefore, when the temperature increases, the oxidation of sediments will be enhanced, that leads to the high hematite content then high redness. This could be why hematite and redness have been widely and successfully used as sensitive proxies in reconstructing climatic changes (e.g., Ji et al., 2002, 2006, 2007; Song et al., 2005; Nie et al., 2010; Liu et al., 2011; Fang et al., 2015).

We also interpreted that the high values of the hematite content and redness throughout the Late Oligocene and Early Miocene strata are records of the strong East Asian monsoon in the region because the higher monsoonal precipitation and temperature is relatively favorable for hematite formation and high redness. Thus, the six short intervals of lower hematite contents and redness values (HR1 to HR6) recorded six short weaker monsoon events with a roughly one million year cycle (Figs. 4 and 7). These events, together with the long-term trend, can be correlated within the age errors, which are caused by using different GPTS times and our sampling intervals, with the global and regional

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**Fig. 7.** Time series of hematite content (a) and redness (b) in cores HZ-1 and HZ-2 based on their magnetostratigraphy (see Fig. 4) and comparison with oxygen isotope records from the South China Sea (c) (Tian et al., 2008) and global marine stacks (d) (Zachos et al., 2008; the heavy line is the 20-point moving average).
temperature records indicated by the global and South China Sea marine oxygen isotope records (Zachos et al., 2008; Tian et al., 2008); events HR1, HR2, HR4 and HR6 are correlated with the Miocene short cooling events Mi-1, Mi-1a, Mi-1aa and Mi-1b (Miller et al., 1987, 1991; Wright et al., 1992), respectively (Fig. 7), and HR3 and HR5 can be correlated with two unrecognized cold events between Mi-1a and Mi-1 and between Mi-1aa and Mi-1b, respectively (Fig. 6). These events occurred on a roughly one million cycle, which appears to have recorded signals of the long-term 0.8-1.2 Ma eccentricity cycles (Milankovitch, 1930; Berger et al., 1984).

Because the hematite content and redness events HR1 to HR6 occur irrespective of lithology (e.g., event HR2 occurs in the sandstone, HR3 occurs in the paleosol, and HR4 and 5 occur in the mudstone), we believe that these events were not affected by tectonic events but rather are real markers of monsoon climate weaknesses caused by global cooling events. Therefore, we conclude that the East Asian monsoon in the Late Oligocene to Early Miocene was principally controlled by both the long-term trend and short-term events of global temperature variations during that period.

6. Conclusions

1) We drilled a new core (HZ-2) that extended 620 m from the bottom of the HLD outcrop section in the southern Linxia foreland basin on the northeastern margin of the Tibetan Plateau and obtained a continuous high quality stratigraphic record that straddles the Late Oligocene to Early Miocene boundary.

2) Detailed paleomagnetic analyses of core HZ-2 yielded an age interval of 21.6-26.5 Ma.

3) The overwhelming occurrence of subaerially oxidized fine sediments in reddish-brown siltstones and mudstones accompanied by many layers of well-developed brownish-red paleosols of luvic cambisols from cores HZ-1 and HZ-2 characterizes a typical floodplain environment and suggests an already occurrence of the East Asian monsoon at least after 26.5 Ma and extended throughout the studied interval of the Late Oligocene to the Early Miocene.

4) The high hematite contents and redness records from cores HZ-2 and HZ-1 confirm the onset of the Late Oligocene - Early Miocene monsoon that was indicated by the lithofacies. They reveal six short intervals of lower values (HR1 to HR6) with a roughly one million year cycle at approximately 22.9 Ma, 21.9 Ma, 21 Ma, 19.7 Ma, 18.9 Ma and 18.1 Ma, respectively, which suggests six short weak monsoon events at those times. Both the long-term trend and the short-term events of the monsoon records correlate well with global and South China Sea marine oxygen isotope records, which suggests that the global temperature was a major forcing for early monsoonal climate change in the Late Oligocene and the Early Miocene at both long and short time scales.

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