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Drastic climatic shift at ~2.8 Ma as recorded in eolian deposits of China and its implications for redefining the Pliocene-Pleistocene boundary

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A B S T R A C T

There has been a long-running debate over the position of the Pliocene-Pleistocene boundary, i.e. 1.8 Ma versus 2.6 Ma. This study presents litho- and pedo-stratigraphic characteristics, grain size, magnetic susceptibility, and dust sedimentation rates for 'red clay'-loess sequences over northern China, in order to assess the drastic climatic event(s) recorded in the quasi-continuous eolian deposits, and to readdress the position of the Pliocene-Pleistocene boundary. In complete Chinese loess sequences, 33 soil-loess couplets have been identified and labeled with *Si-Li* system. Traditionally, the base of Chinese loess has been designated at the bottom of loess unit L33. However, a typical loess layer within the uppermost 'red clay' is identified and formally designated here as L34 and the soil unit between L33 and L34 as S33, which lowers the basal age of Chinese loess from 2.6 Ma to ~2.8 Ma. Although the Neogene 'red clay' and the overlying Quaternary loess are both of windblown origin, they have quite different pedogenic characteristics. The loess consists of alternating reddish soil and yellowish loess layers, while the 'red clay' is composed of soils with much redder colour and carbonate nodule layers. In general, the "red clay" has been subjected to stronger pedogenic processes than the overlying loess, and can be regarded as an extremely thick soil complex. The alternation of loess and soils reflects large-scale oscillations between glacial and interglacial conditions, whereas the well-developed 'red clay' deposition indicates a relatively stable climate condition. At ~2.8 Ma, shortly before the Gauss/Matuyama boundary, the eolian sequences show a sedimentary transition from the well-weathered 'red clay' to the alternating loess and paleosol deposits, accompanied by coarser particle size, increased dust accumulation rates, and change of magnetic susceptibility signature. These indicate a large climatic shift from long-lasting warm-humid conditions to large-amplitude cold-dry and warm-humid fluctuations around 2.8 Ma, and a dramatic increase in aridity over the dust source region. In contrast, such a drastic event is not identified around the top of the Olduvai event or at any other time of the past 7.7 Ma. As the Gauss/Matuyama boundary offers a sound foundation for global time correlation and chronostratigraphic classification, these results still support the proposal of ICS and INQUA, that the conventional Pliocene-Pleistocene boundary should be lowered from 1.8 Ma to 2.6 Ma.

1. Introduction

In 1984, the Pliocene-Pleistocene boundary was defined by a Global Boundary-Stratotype Section and Point (GSSP) at Vrica in Calabria, southern Italy, at the 27th International Geological Congress in Moscow. In the Vrica section, the base of the marine claystones conformably overlying the sapropelic bed '*e*', near the top of the Olduvai event (~1.8 Ma), was selected as the Pliocene-Pleistocene boundary marker point or 'golden spike' (Tauxe et al., 1983; Aguirre and Pasini, 1985; Zijderveld et al., 1991; Nikiforova and Alekseev, 1997; Pasini and Colalongo, 1997). The primary reason for this decision is that several micropaleontological events around the proposed boundary-stratotype are approximately coeval with the beginning of a cold-climate phase marked by the first appearance of a so-called 'northern guest', *Arctica islandica*, in Italian shallow-water sections. The appearance of this famous 'northern guest' in the Mediterranean has long been one of the main criteria for defining the base of the Pleistocene. However, considerable debate has arisen concerning the designated Pliocene-Pleistocene boundary ever since (e.g. An, 1984; Liu, 1985; Jenkins, 1987; Wang et al., 1991; Bardají et al., 1997; Ding et al., 1997; Naish et al., 1997; Partridge, 1997; Suc et al., 1997; Vai, 1997; Zhang, 1997; Mauz, 1998; Wang, 2000; Gibbard et al., 2005). Increasing evidence has shown that major cooling events in the Mediterranean

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occurred around 2.6–3.0 Ma (e.g. Versteegh, 1997; Monegatti and Raffi, 2001; Roveri and Taviani, 2003), and that the final definition of the Vrica Pliocene-Pleistocene boundary was based neither on bio-events, nor climatic criteria (e.g. Jenkins, 1987; Partridge, 1997; Suc et al., 1997). Thus, lowering the Pliocene-Pleistocene boundary from the top of the Olduvai subchron to the Gauss/Matuyama magnetic reversal has been strongly proposed.

Following the controversy over the Pliocene-Pleistocene boundary, the term 'Quaternary' has recently come under review, with proposals for its abandonment or modification (e.g. Gradstein et al., 2004; Liu, 2004; Ogg, 2004; Pillans, 2004; Pillans and Naish, 2004; An and Ai, 2005; Aubry et al., 2005; Gibbard et al., 2005; Suguio et al., 2005). In 2007, after long debate, the International Commission on Stratigraphy (ICS) and the International Union for Quaternary Research (INQUA) jointly proposed that the Quaternary be established as a System of the Cenozoic Erathem, with its base defined by the GSSP of the Gelasian Stage, and that the base of the Pleistocene be lowered to coincide with the Gelasian Stage GSSP at ~2.6 Ma to maintain strict hierarchy in the Geologic Time Scale (Head et al., 2008; Ogg and Pillans, 2008). This proposal was ratified by the International Union of Geological Sciences (IUGS) in June 2009.

The conventional Pliocene-Pleistocene boundary-stratotype was established in marine sediments. As a GSSP, the Pliocene-Pleistocene boundary should have global significance and should be documented in well-preserved continental records (Salvador, 1994). Thus, the Pliocene-Pleistocene boundary should also be clearly identified in relatively continuous continental records. Loess deposits in northern China are one of few quasi-continuous terrestrial sediments covering the whole period of the Quaternary (Liu, 1985; An et al., 1990; Liu and Ding, 1998; Ding et al., 2002). Chinese loess is well correlated with the marine isotopic record (Kukla, 1987; Shackleton et al., 1995; Ding et al., 2002), and hence is a valuable archive for regional climate and for assessing the position of the Pliocene-Pleistocene boundary.

The base of Chinese loess has been dated around the Gauss/Matuyama magnetic reversal (Heller and Liu, 1982; Liu, 1985; Ding et al., 1997). In the Loess Plateau, complete loess-soil sequences are generally underlain by reddish clay-silt-sized sediments, informally known as the 'red clay' formation (Liu, 1985). Field observations and geochemical and sedimentological studies have demonstrated the windblown origin for the 'red clay', thus extending the relatively continuous eolian record in the Chinese Loess Plateau back in time from 2.6 to 7.0–7.7 Ma (e.g. Zheng et al., 1992; Ding et al., 1998, 1999a, 2001a; Sun et al., 1998a; Han et al., 2002; Yang and Ding, 2004), the oldest eolian deposits dated back even to 22 Ma (Guo et al., 2002).

On the basis of the connotation of dry, cold-climate conditions for Chinese loess, Tungsheng Liu, Zhisheng An, and their colleagues have proposed that the onset of extensive loess deposition over northern China marks the beginning of the Pleistocene, and that the Gauss/Matuyama reversal is more appropriate to be the Pliocene-Pleistocene boundary (Liu and Ding, 1982; An, 1984; Liu, 1985; Liu et al., 1985). These contentions have been supported by subsequent studies (Kukla, 1987; Ding et al., 1997). This study presents litho- and pedo-stratigraphic characteristics, grain size, magnetic susceptibility, and dust sedimentation rates from 'red clay'-loess sequences at Jingchuan, Lingtai, Baoji, and Lantian (Fig. 1) on the Chinese Loess Plateau. The aim is to assess the drastic climatic event(s) recorded in the eolian deposits and to readdress the position of the Pliocene-Pleistocene boundary.

2. Site descriptions

The Lingtai (35.01°N, 107.51°E) and Jingchuan (35.29°N, 107.37°E) sections are situated in the central part of the Loess Plateau (Fig. 1), with mean annual temperature and precipitation values of 9–10 °C and 550–600 mm respectively. The Lingtai 'red clay'-loess sequence has a thickness of 305 m and consists of 175 m of loess and 130 m of 'red clay'. The Jingchuan section, with a thickness of 325 m, is composed of 199 m of loess and 126 m of 'red clay'. Palaeomagnetic studies (Ding et al., 1998, 2001a) suggest a basal age of 7.05 Ma for the Lingtai section and 7.7 Ma for the Jingchuan section (Fig. 2).

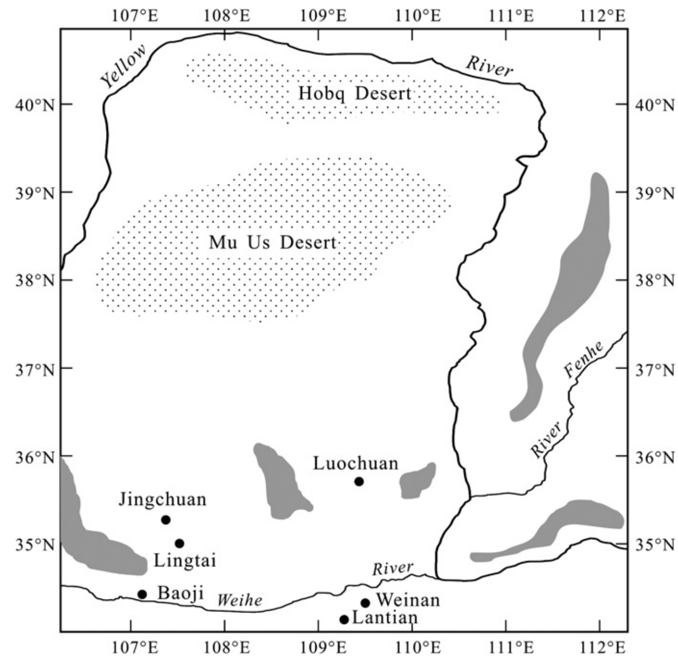

Fig. 1. Schematic map showing the study localities. Also shown are the deserts (dotted) and mountains (grey) around and within the Loess Plateau.

The Baoji (34.43°N, 107.12°E) and Lantian (34.19°N, 109.24°E) sections are located in the southernmost part of the Plateau (Fig. 1), with mean annual temperature and precipitation values of 13 °C and 670–700 mm respectively. The Baoji section has a thickness of 187 m and consists of 160 m of loess and 27 m of 'red clay' (Fig. 2; Evans et al., 1991; Rutter et al., 1991). The Lantian section, about 195 m thick, is composed of 136 m of loess and 59 m of 'red clay' (Zheng et al., 1992). The Baoji and Lantian sections have a basal age of ~4.4 Ma (Fig. 2; Evans et al., 1991) and ~5.2 Ma (Fig. 2; Zheng et al., 1992), respectively, recalculated using the time scale of Cande and Kent (1995).

For all the sections, a total of 21,990 samples were collected at 3–5 cm intervals. Bulk magnetic susceptibility and grain size were measured for all samples with a Bartington MS2 susceptibility meter and a SALD-3001 laser diffraction particle analyser. The particle analytical procedures were as detailed by Ding et al. (1999b).

3. Drastic climatic event(s) recorded in the 'red clay'-loess sequences

3.1. Litho- and pedo-stratigraphic characteristics

In general, loess deposits consist of two major stratigraphic units, loess and paleosol. Loess horizons are labeled *Li* and soils *Si*. A striking feature of the loess is that it consists of alternating reddish soil and yellowish loess layers (Fig. 3). On the Chinese Loess Plateau, 33 soil-loess couplets have been identified in complete loess sequences (Rutter et al., 1991; Ding et al., 1993). Loess units, with

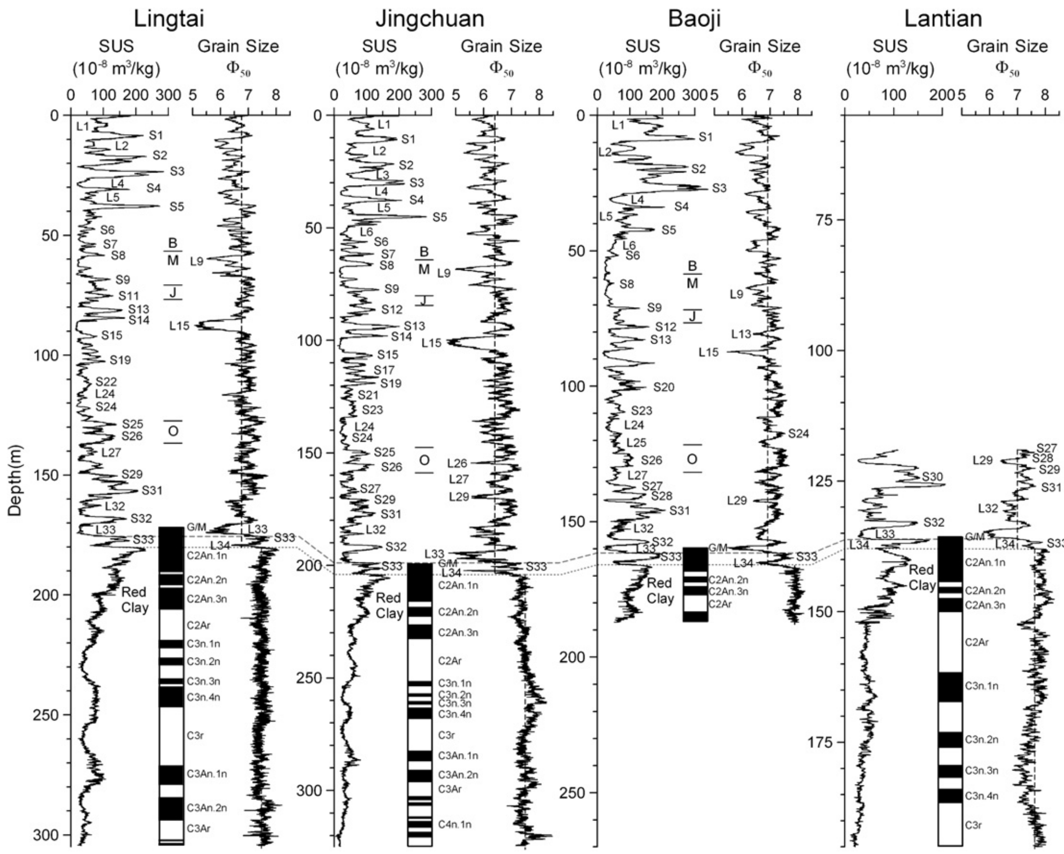

Fig. 2. Magnetic susceptibility (SUS) and median grain size (Φ_{50}, phi scale) records of the 'red clay'-loess sequences at Lingtai, Jingchuan, Baoji, and Lantian, together with the magnetic reversal polarities. Most of the soil units (*Si*) and major loess beds (*Li*) are indicated. The base of Chinese loess has been traditionally designated at the bottom of loess unit L33 (horizontal dashed lines), and is lowered to the bottom of L34 (horizontal dotted lines) in this study. The 'red clay'-loess contact is assigned a paleomagnetic age of ~2.8 Ma, shortly before the Gauss/Matuyama (G/M) reversal. The vertical dashed lines indicate the averaged values of median grain size in the 'red clay' and the overlying loess. The magnetic polarity of Baoji is modified from Evans et al. (1991) and Rutter et al. (1991), and that of Lantian from Zheng et al. (1992).

varying thickness, are yellowish, massive, and have high carbonate contents, whereas soil units exhibit a brownish or reddish colour and obvious pedogenic features, such as organic matter accumulation, sponge or subangular block structures, clay and/or Fe–Mn skins, and carbonate pseudomycelia and/or nodules.

Traditionally, the loess sequence has been divided into three formations—the Wucheng (S15-L33, ~1.3–2.6 Ma), Lishi (S1-L15, ~0.07–1.3 Ma), and Malan (L1, ~0.01–0.07 Ma) (Liu, 1985; Kukla, 1987; Rutter et al., 1991; Ding et al., 1993, 2002). The Wucheng Loess is separated from the overlying Lishi Loess by stratigaphic marker L15, a unit of thick, coarse silt readily recognizable in the field. The Lishi Loess is overlain by the Malan Loess (L1), which accumulated during the last glacial and supports the Holocene soil (S0). The base of Chinese loess has been traditionally designated at the bottom of loess unit L33 (Rutter et al., 1991; Ding et al., 1993; Liu and Ding, 1998). On the Loess Plateau, however, a 1- to 4-m-thick yellowish and massive loess layer, with coarse gain size and low magnetic susceptibility, is clearly seen in the uppermost 'red clay' in several representative sections (Figs. 2 and 3; Ding et al., 1999a; Han et al., 2002). It is formally designated as L34, and the soil unit between L33 and L34 as S33. In the Lingtai and Jingchuan sections, paleosol S33 consists of two discrete soils and a thin intervening loess horizon, while no loess horizon within S33 can be recognized at Baoji and Lantian. Lithologically, the S33–L34 couplet can be assigned to the Wucheng Loess. Thus the basal age of Chinese loess is lowered from 2.6 Ma (L33) to ~2.8 Ma (L34).

The most striking features of the 'red clay' are its reddish colour and the alternation of soils and carbonate nodule layers (Fig. 3C). For example, over 110 couplets of pedogenic B horizons and horizontal carbonate nodule horizons are identified in the Lingtai 'red clay' sequence (Ding et al., 1999a). In general, the 'red clay' has been subjected to stronger pedogenic processes than the overlying loess, and can be regarded as an extremely thick soil complex since no typical loess horizons are seen within it (Fig. 3C).

Paleosols within loess-'red clay' sequences are accretionary in nature as dust deposition continued throughout soil-forming intervals, which may lead to upward migration of soil genetic horizons (Bronger and Heinkele, 1989; Guo et al., 1991; Kemp et al., 1995, 1996; Han et al., 2002). Moreover, paleosols have undergone a number of diagenetic changes including loss of organic matter, burial reddening due to aging of hydrous ferric oxides, and compaction (e.g. Retallack, 1991). In this context, buried paleosols are somewhat different from modern soils and can be only roughly correlated with soil maps that depict modern surface soils (Soil Survey Staff, 1999).

According to the description of the pedo-stratigraphic units for the Lingtai section (Ding et al., 1999a), the strongly developed soil horizons in the 'red clay' are dark reddish brown (2.5YR), and have strong fine subangular blocky structure, many thick translocated clay skins, and many Fe–Mn films. These soils approximately correspond to Udalfs in the Alfisol order (Soil Survey Staff, 1999). The weakly developed 'red clay', with reddish brown (5YR) colour, weak coarse subangular blocky structure, and few clay skins, is similar to Ustalfs in the Alfisol order (Soil Survey Staff, 1999). The Pleistocene loess units (yellowish brown (10YR); massive) appear to be non-weathered or only slightly weathered, and can be

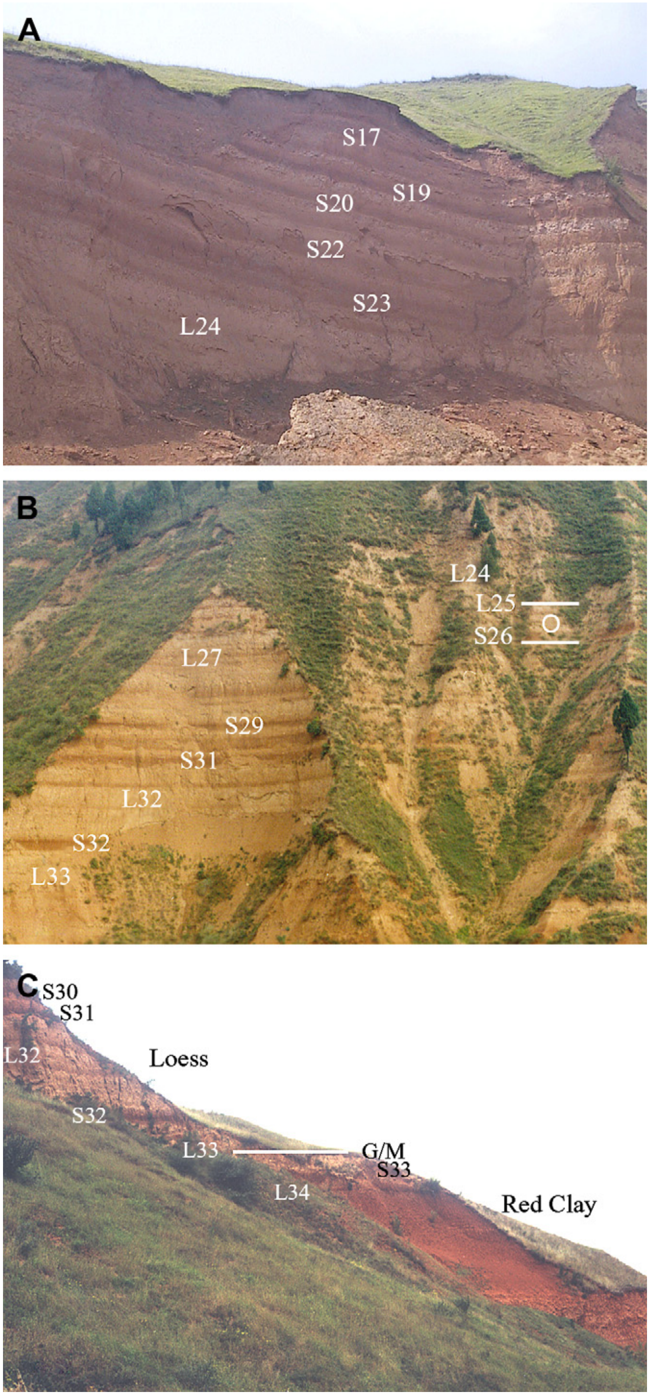

Fig. 3. Photographs showing the eolian deposits from the Chinese Loess Plateau. (A) the S17–L24 portion at Lingtai. (B) the L24–L33 portion at Weinan (Fig. 1). (C) the S30–L34 portion of loess and the upper part of 'red clay' at Lantian (courtesy of J.T. Han). The top and base of the Olduvai (O) subchron and the Gauss/Matuyama (G/M) boundary are indicated by the white lines.

assigned to Aridisols (Soil Survey Staff, 1999). The strongly developed Pleistocene soils (S4, S5, S8, S13, S26, S31, and S33) are dark reddish brown (5YR), and have strong coarse to fine prismatic structure or moderate medium subangular blocky structure and some to many clay skins. These soils can be approximately classified as Udalfs. The other soils within loess are generally reddish brown (5YR to 7.5YR), and have weak to moderate subangular blocky structure and few to common clay skins. These weakly developed soils are roughly equivalent to Ustalfs.

Paleomagnetic studies from several loess sections have shown a generally consistent result for the positions of major geomagnetic reversal boundaries (Fig. 2; Rutter et al., 1991; Ding et al., 1998, 2001a, 2002). The Matuyama/Brunhes geomagnetic reversal is located in loess unit L8, and the upper and lower reversals of the Jaramillo subchron are defined, respectively, within L10 and L12. The Olduvai subchron is detected between the middle part of L25 and the lower part of S26 or the uppermost part of L27 (Figs. 2 and 3B). The Gauss/Matuyama reversal occurs within loess unit L33 (Figs. 2 and 3C). The conventional Pliocene-Pleistocene boundary is located in loess unit L25 (Fig. 3B). However, field observations have shown that there is no litho- and pedo-stratigraphic contrast between the loess deposits below L25 (Fig. 3B) and those above (Fig. 3A), with both exhibiting regular loess-soil alternations. In contrast, a striking difference is observed shortly before the Gauss/Matuyama boundary at ~2.8 Ma, which corresponds to the 'red clay'-loess contact (Fig. 3C).

3.2. Grain size

The alternation of loess and soils is clearly expressed in the grain-size curves (Fig. 2). Throughout sections, loess horizons show coarse grain-size values, while paleosols are characterized by fine particle sizes. In the Lingtai section, the median grain size falls in the range 5.0–6.6 Φ for loess units, and 6.6–7.5 Φ for soils. The values are 4.5–6.5 Φ for loess units and 6.5–7.3 Φ for soils at Jingchuan, and 5.5–7.0 Φ for loess units and 6.8–7.5 Φ for soils at Baoji. In the Lantian section, the median grain size values are 5.8–7.0 Φ for loess units and 7.2–7.7 Φ for soil units. The 'red clay' at the four sites all show fine size values, centered around 7.2–8.0 Φ. Clearly, the grain size shows no striking changes between the loess portions below and above the Olduvai event, with all exhibiting regular oscillations. In contrast, at ~2.8 Ma, there is a dramatic coarsening shift of the grain size.

3.3. Magnetic susceptibility

The alternation of loess and soils is also clearly expressed in the magnetic susceptibility records (Fig. 2). Paleosols are characterized consistently by higher susceptibility values (100–300 $\times$ 10^{-8} m^3/kg) compared with the loess horizons (<70 $\times$ 10^{-8} m^3/kg) above and below them. In the four sections, the susceptibility values of 'red clay' fall in the range 10–100 $\times$ 10^{-8} m^3/kg except the late Pliocene portion (100–220 $\times$ 10^{-8} m^3/kg). Two differences in magnetic susceptibility between the 'red clay' and the overlying loess are observed. First, the 'red clay' generally shows much lower susceptibility values than the overlying Quaternary paleosols, despite its higher pedogenic degree. Second, the magnetic susceptibility exhibits low-frequency, small-amplitude fluctuations in the 'red clay' and high-frequency, large-amplitude variations in the loess-soil sequences. Clearly, the striking change of susceptibility signature is observed near the Gauss/Matuyama boundary rather than around the Olduvai event.

3.4. Eolian sedimentation rates

In the four sections, the sedimentation rates are generally low (~10–35 m/Ma) for the late Miocene and Pliocene, and show a stepwise increase since 2.6 Ma (Fig. 4). In the Jingchuan section, the sedimentation rate increased abruptly from ~37 m/Ma to ~64 m/Ma at 2.6 Ma, and further to ~82 m/Ma after 1.8 Ma. At Lingtai, it rose from ~40 m/Ma to ~56 m/Ma at 2.6 Ma, and reached ~70 m/Ma after 1.8 Ma. In the Baoji section, it increased from ~19 m/Ma to ~44 m/Ma at 2.6 Ma, and further to over 64 m/Ma after 1.8 Ma. At Lantian, it rose from ~15 m/Ma to ~40 m/Ma at

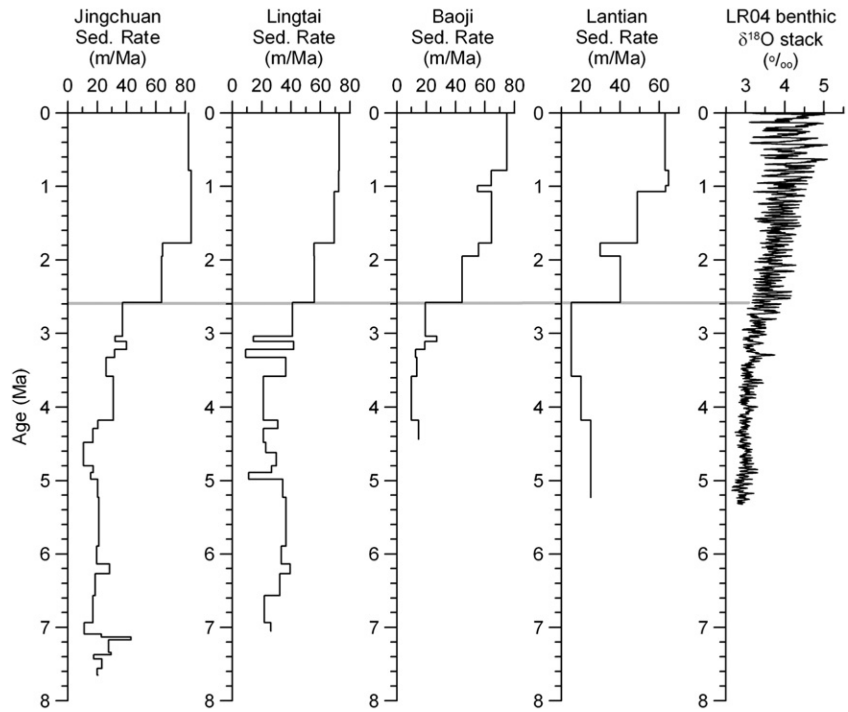

Fig. 4. Linear sedimentation-rate models for the 'red clay'-loess sequences at Jingchuan, Lingtai, Baoji, and Lantian, and correlation with a stacked benthic $\delta^{18}O$ record (Lisiecki and Raymo, 2005). The sedimentation rate of Lantian is calculated using the data from Zheng et al. (1992). Ages of the four eolian sequences are constrained by magnetic polarity reversals according to the Cande and Kent (1995) geomagnetic polarity timescale. The abrupt increase in sedimentation rate at 2.6 Ma (Gauss/Matuyama boundary) is marked by the grey horizontal line.

2.6 Ma, then increased to ~49 m/Ma for the time interval 1.8–1.1 Ma, and further to ~63 m/Ma after 1.1 Ma.

4. Discussion

The alternation of loess and soils documents large-scale oscillations between glacial and interglacial conditions (Liu, 1985; Kukla, 1987; Rutter et al., 1991; Ding et al., 1993). During glacial periods, desert environments expanded southeastward due to a weak East-Asian summer monsoon, leading to coarse particle size and high sedimentation rate for loess units. In interglacial periods, desert environments retreated northwestward as the summer monsoon strengthened, resulting in the development of soils with fine particle size and low sedimentation rate (Ding et al., 2001b; Yang and Ding, 2008). Therefore, changes in grain size and sedimentation rate of Chinese loess are closely related to the aridity conditions of inland China.

In loess-soil sequences, the magnetic susceptibility is roughly positively correlated with the pedogenic degree and shows wide variations, although the processes and factors affecting it remain controversial (e.g. Kukla et al., 1988; An et al., 1990; Zhou et al., 1990; Maher and Thompson, 1992; Lu et al., 2000; Sun and Liu, 2000; Chen et al., 2005; Liu et al., 2007; Torrent et al., 2007). The susceptibility of the 'red clay' is lower overall than the paleosols in the overlying loess, although the pedogenic development of the 'red clay' is apparently better than that of the soils in the loess. For instance, the soils between 220 m and 240 m (4.0–4.8 Ma) at Lingtai have many thick clay skins and common Fe–Mn films, indicating high monsoonal rainfall during their development (Ding et al., 1999a), but the susceptibility values in this interval are as low as the values in the non-weathered loess layers of Pleistocene age. This case has also been observed in the Miocene 'red clay' from the western Loess Plateau (Hao et al., 2008).

The difference in susceptibility signature between the 'red clay' and the overlying loess-soil sequence may be explained by one or more of the following reasons. First, the overall reduction of magnetic susceptibility in the 'red clay' may be inherited from the dust source material (Ding et al., 2001c). Second, the magnetic minerals in the 'red clay' could have been destroyed by gleying process, as indicated by abundant dark Fe–Mn films on the soil structural surfaces (Ding et al., 1999a, 2001c). Third, pedogenic maghemite and hematite form through the same pathway (ferrihydrite → maghemite → hematite) (Torrent et al., 2007), i.e. the formation of hematite at the expense of maghemite under intense weathering conditions have resulted in the low susceptibility values of the 'red clay'. Although the paleoclimatic significance of the magnetic susceptibility signal within the 'red clay' remains controversial (e.g. Sun et al., 1998b; Ding et al., 1999a, 2001c; An, 2000; Liu et al., 2003; Hao et al., 2008), a substantial change in climate conditions and/or dust source regions at ~2.8 Ma is responsible for the difference in susceptibility signature between the 'red clay' and the overlying loess-soil sequence.

The first deterioration of climate over northern China occurred at ~2.8 Ma (L34). This event is well represented by the transition from Alfisols to alternating Aridisols and Alfisols above, accompanied by coarser particle size, increased dust accumulation rates, and change of magnetic susceptibility signature. These indicate a large climatic shift from long-lasting warm-humid conditions to large-amplitude cold-dry and warm-humid fluctuations around 2.8 Ma, and a dramatic increase in aridity over the dust source region of northern China. Pollen records from the central and southern Loess Plateau have shown a striking climatic deterioration at 2.7–2.6 Ma (Han et al., 1997; Wu et al., 2007), which is consistent with the conclusion from litho- and pedo-stratigraphic features, grain size, magnetic susceptibility, and sedimentation rates of the eolian sequences.

In contrast, there are no notable changes for loess deposits around 1.8 Ma except for their sedimentation rates. Before and after 1.8 Ma, the loess stratigraphy shows a periodic loess-soil alternation, and the grain size and magnetic susceptibility exhibit regular oscillations. The dust accumulation rate shows a 1.4- to 2.7-fold increase at 2.6 Ma, while only a 1.2- to 1.5-fold increase is observed at 1.8 Ma. In this context, a large climatic shift occurred shortly before the Gauss/Matuyama magnetic reversal at ~2.8 Ma, whereas no such an event is identified around the top of the Olduvai subchron or at any other time of the past 7.7 Ma, though a further drier climate after 1.8 Ma is inferred from the sedimentation-rate changes.

Marine records have shown a major phase of Pliocene northern hemisphere ice growth between 3.1 Ma and 2.7 Ma (Fig. 4; Tiedemann et al., 1994; Lisiecki and Raymo, 2005), and a dramatic increase in ice-rafting debris in the northwest Pacific and the Norwegian Sea at ~2.8 Ma (Maslin et al., 1995). The abrupt climatic shift around 2.8 Ma in northern China, marked by the onset of extensive loess deposition, indicates a rapid reorganization of regional climate system in East Asia, which may be causally related to the initiation of extensive northern hemisphere glaciation (Ding et al., 2000; Han et al., 1997, 2002; Xiong et al., 2003). The climatic deterioration around 2.6–2.8 Ma is not only well recorded in Chinese loess-'red clay' sequence (e.g. Ding et al., 1992, 1997, 2000; Liu and Ding, 1993; Han et al., 2002; Xiao and An, 1999; An, 2000), but is also clearly documented in the records of continental mammal faunas in East Asia (Liu and Ding, 1982; Zhang, 1997), West Europe (Azzaroli et al., 1997) and Africa (Cooke, 1997), and in other records on the globe (e.g. Zacwijn, 1974; Song et al., 1982; Thouveny and Servant, 1989; Zachariasse et al., 1990; Naish et al., 1997; Han et al., 1997). It is thus a drastic environmental event of global significance.

Since the pioneering work of Liu and his collaborators (Liu and Ding, 1982; An, 1984; Liu, 1985; Liu et al., 1985), the onset of extensive loess deposition has been regarded to mark the beginning of the Pleistocene. On the Loess Plateau, a typical loess unit, formally labeled as L34, is identified within the uppermost 'red clay', which allows us to lower the base of Chinese loess from 2.6 Ma to ~2.8 Ma. In this regard, the Pliocene-Pleistocene boundary may be placed at 2.8 Ma, some 200 ka earlier than previously thought by the Chinese loess community (e.g. An, 1984; Liu and Ding, 1982; Liu, 1985, 2004; Liu et al., 1985; Kukla, 1987; Rutter et al., 1991; Ding et al., 1997; An and Ai, 2005). As magnetic polarity reversals offer a sound foundation for global time correlation and chronostratigraphic classification, these results still support the choice of 2.6 Ma as the Pliocene-Pleistocene boundary.

The terms 'Pleistocene' and 'Quaternary' have long been endowed with a connotation of glaciation and are both in widespread use within and well beyond the geological community (Forbes, 1846; King and Oakley, 1949; Liu, 2004; Pillans and Naish, 2004; An and Ai, 2005; Gibbard et al., 2005). According to the International Stratigraphic Guide (Salvador, 1994), the selection of the boundary-stratotypes of chronostratigraphic units should take account of historical priority and usage, and should favor long-distance time correlation. As reviewed by Suc et al. (1997) and Partridge (1997), the Vrica Pliocene-Pleistocene boundary is very weak and the Monte San Nicola section (Rio et al., 1998) in southern Sicily might replace the Vrica section as the Pliocene-Pleistocene boundary-stratotype section. Clearly, lowering the present Pliocene-Pleistocene boundary to the Gauss/Matuyama magnetic reversal and retaining the Quaternary as a formal international Period/System with its base at 2.6 Ma best meets the requirements of boundary-stratotypes of chronostratigraphic units.

5. Conclusions

In northern China, extensive loess deposition began at ~2.8 Ma, some 200 ka earlier than previously thought. At ~2.8 Ma, the eolian sequences show a sedimentary transition from the well-weathered 'red clay' to the alternating loess and paleosol deposits, accompanied by coarser particle size, increased dust accumulation rates, and change of magnetic susceptibility signature. These indicate a large climatic shift from long-lasting warm-humid conditions to large-amplitude cold-dry and warm-humid fluctuations around 2.8 Ma, and a dramatic increase in aridity over the dust source region. In contrast, such a dramatic event is not identified around the top of the Olduvai subchron or at any other time of the past 7.7 Ma. As the Gauss/Matuyama magnetic reversal is very important for global time correlation and chronostratigraphic classification, the eolian records from China support the lowering of the conventional Pliocene-Pleistocene boundary from 1.8 Ma to 2.6 Ma.

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