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ScienceDirect
Physics of the Earth and Planetary Interiors 159 (2006) 109–117
PHYSICS OF THE EARTH AND PLANETARY INTERIORS
www.elsevier.com/locate/pepi

A magnetostratigraphic reassessment of correlation between Chinese loess and marine oxygen isotope records over the last 1.1 Ma

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Received 7 September 2005; received in revised form 22 June 2006; accepted 12 July 2006

Abstract

We present a magnetic polarity-based paleoclimatic correlation between Chinese loess and marine oxygen isotope stage (MIS) records over the last 1.1 Ma. This revised land–ocean comparison indicates that inconsistent occurrence of the Matuyama/Brunhes Boundary, which lies in MIS 19 in marine sediments and occurs either at the base/lower part of loess L8 or the uppermost part of soil S8 in Chinese loess, is due to regional and/or local climate variability across the Chinese Loess Plateau, and does not need to be attributed to lock-in effects of loess remanence acquisition associated with pedogenesis. This result implies that the geomagnetic polarity boundaries in Chinese loess in general are reliable chronostratigraphic time markers. Therefore astronomically tuned timescales for Chinese loess that presume a downward shift of polarity boundaries due to a time lag in remanence acquisition may have overestimated ages of certain loess/paleosol units.

Keywords: Loess; Magnetostratigraphy; Marine sediments; Lock-in; Paleoclimate; Magnetic susceptibility

1. Introduction

Both marine sediments and Chinese loess carry essentially continuous records of geomagnetic polarity reversals and paleoclimate that can be serially correlated on a global scale (e.g., Kukla et al., 1988; Heller and Evans, 1995; Zhu et al., 1998a; Evans and Heller, 2001; Ding et al., 2002). In marine sediments, the climate proxy is based on the variations of oxygen isotope ratios ($^{18}O/^{16}O$) in benthic foraminifera that increase during colder periods (glacials) when global ice volume increases, and vice versa (Ruddiman et al., 1989; Raymo et al., 1989; Shackleton et al., 1990). The alternating loess–paleosol sequences on the Chinese Loess Plateau (CLP) are associated with the dominance of the winter monsoon that blows from Siberia during cold climate periods, and the southeasterly warm–moist summer monsoon, respectively (Liu, 1985; Liu and Ding, 1998; An, 2000). While loess has relatively low values of magnetic susceptibility (MS), interbedded paleosols formed by pedogenic processes during the summer monsoon conditions have larger MS values (Kukla et al., 1988; Zhou et al., 1990). There is a compelling resemblance between $\delta^{18}O$ records of foraminifera in marine sediments and MS signal on the CLP, indicative of a close linkage between aeolian flux, global ice volume and climate (Ding et al., 1994, 2002; Heller and Evans, 1995; Liu and Ding, 1998).

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doi:10.1016/j.pepi.2006.07.002

However, a straightforward correlation between the loess MS and oxygen isotope profiles results in a major discrepancy between the positions of the Matuyama–Brunhes Boundary (MBB) in marine sediments and Chinese loess sections. While the MBB in marine sediments is well established within interglacial marine isotope stage (MIS) 19 (778.0 ± 1.7 ka (Tauxe et al., 1996)), it is commonly observed within the glacial loess layer L8 and occasionally in the interglacial paleosol S8 on the CLP (Heller and Liu, 1982; Zhu et al., 1994, 1998a; Pan et al., 2002; Kukla and An, 1989; Rutter et al., 1991). Assuming a paleoclimatic consistency between the stratigraphy of loess and marine sediments, most studies propose that paleosol S7 should be correlated with MIS 19 (Heller and Evans, 1995; Zhou and Shackleton, 1999). Consequently, the MBB appears to be recorded earlier in Chinese loess than in marine sediments (Tauxe et al., 1996; Heslop et al., 2000; Ding et al., 2002). This MBB discrepancy has traditionally been explained by a varying time lag in the magnetization acquisition process in loess causing the observed downward displacement of specific polarity boundaries (Zhou and Shackleton, 1999). This may be accounted for if a detrital remanent magnetization (DRM) becomes locked-in some time after deposition, or if pedogenic processes produce a chemical remanent magnetization (CRM) at some depth below the accumulating surface. Both processes will cause a downward shift in position of a polarity boundary (Zhou and Shackleton, 1999; Spassov et al., 2003).

In order to clarify the fidelity of reversal boundaries and extent of lock-in depth of remanence acquisition in Chinese loess, we present a correlation between detailed magnetostratigraphic records from a loess section at the SE extremity of the CLP and several well-documented MIS records from the North Atlantic. The objective is to assess the significance of downward shifts of paleomagnetic signatures with respect to lateral climate variations across the CLP.

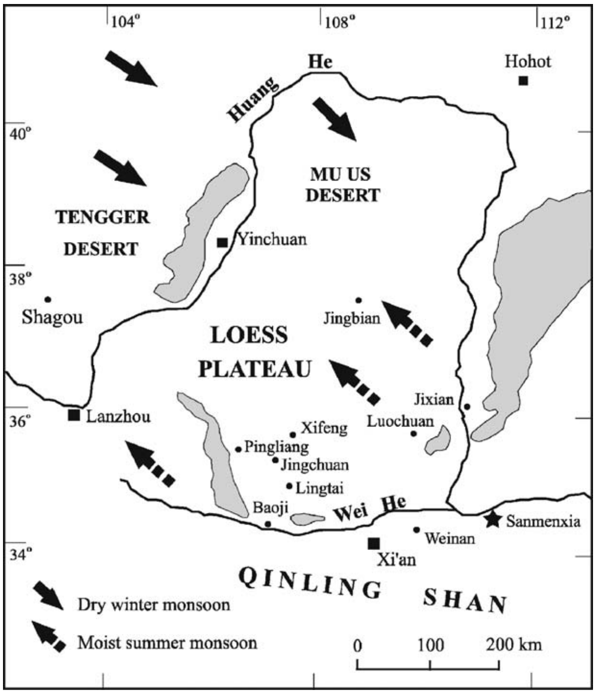

Fig. 1. Map showing the Chinese Loess Plateau (CLP). The Sanmenxia section is marked with a star, and the mountains along and within the CLP are indicated with shaded zones.

2. Paleomagnetic sampling and measurements

The Sanmenxia loess section (34°38′N, 111°09′E) is located at the southeastern margin of the CLP, 15 km southwest of Shaanxian County, Henan Province (Fig. 1). The 153-m-thick section consists of 145-m-thick loess–paleosol sequences overlying 8-m-thick red clay deposits. Oriented samples were collected at 10 cm interval from the top of L1 to the upper part of L13 excluding the topmost 1.4 m because of agricultural activity. From L8 to the upper part of S8 and from the bottom of S9 to L13, samples were continuously collected to get the complete MBB and Jaramillo Normal Subchron (JNS). In the laboratory, every independently oriented sample collected at each level was then cut into three 2 cm × 2 cm × 2 cm specimens (named sets A, B and C). Natural remanent magnetization (NRM) and demagnetization measurements were made on a 2G-755 cryogenic magnetometer. All set A specimens were subjected to progressive thermal demagnetization (20–50 °C intervals) to 580 °C or 680 °C using an ASC-TD48 thermal demagnetizer. For all set B specimens, low-field MS was determined on a Bartington MS2B. Saturation isothermal remanent magnetization (SIRM) was imparted in a DC field of 2 T, and then a 0.1 T reversed-field IRM was imposed to enable calculation of the ratio $S_{-0.1\,\mathrm{T}} = \mathrm{IRM}_{-0.1\,\mathrm{T}}/\mathrm{SIRM}$.

3. Paleomagnetic results

The paleomagnetic results from Sanmenxia have previously been published (Wang et al., 2005), and only a summary will be given here. A secondary normal-polarity magnetization is generally erased below 300 °C (Fig. 2). The characteristic remanent magnetization (ChRM) was typically isolated between 350 °C and 550 °C. The MBB lies at the top of S8 (Fig. 3), in accord

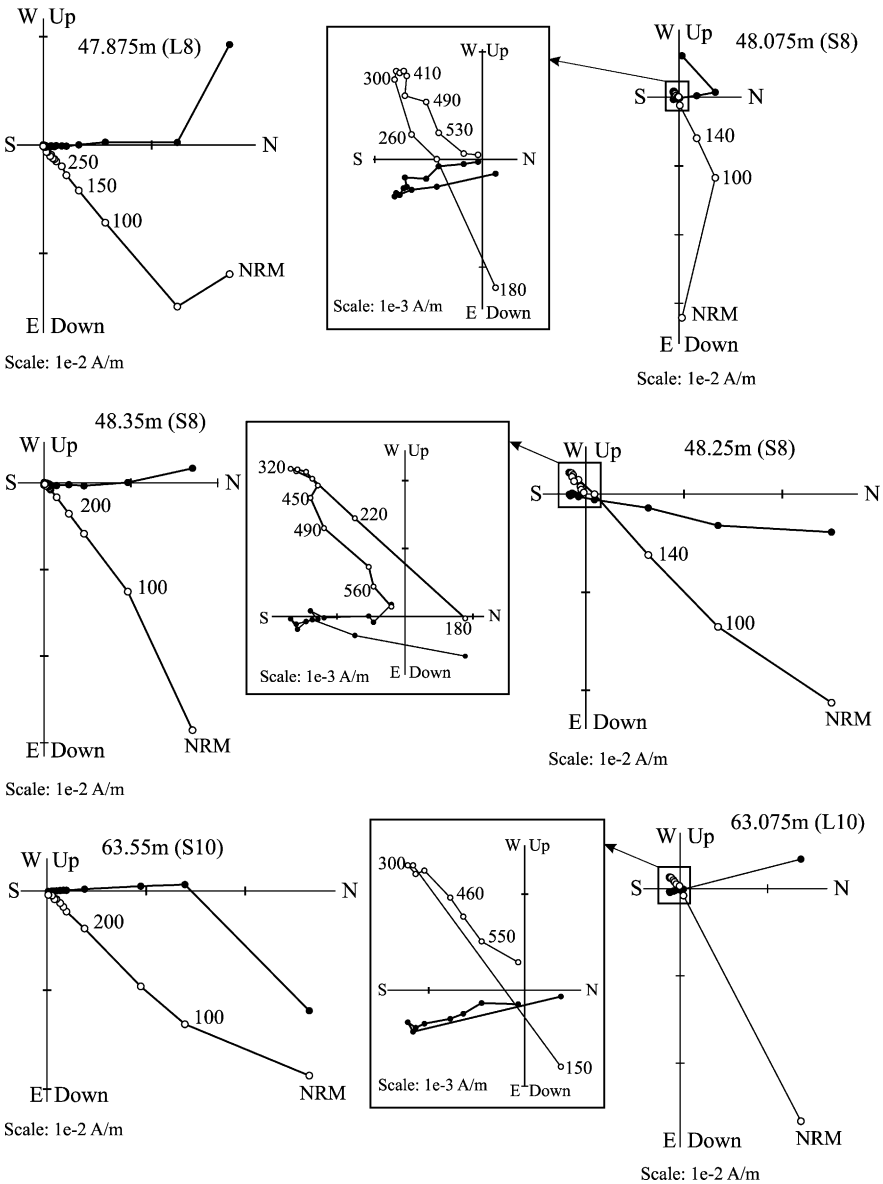

Fig. 2. Orthogonal projections of progressive thermal demagnetization data for the samples across the Matuyama–Brunhes boundary and Jaramillo normal subchron. Solid (open) circles refer to the projection on the horizontal (vertical) plane. The depth of each sample is indicated.

with previous results from Luochuan, Jixian and Fanshan (Heller and Liu, 1982; Hus and Han, 1992; Xiong et al., 2001), but different from results from Weinan (Zhu et al., 1994; Pan et al., 2002), Xifeng (Liu et al., 1987) and Pingliang (Sun et al., 1998) (MBB: base of L8), Shagou (Pan et al., 2001; Wu et al., 2005), Jingbian (Ding et al., 1999; Guo et al., 2002), Baoji (Rutter et al., 1991; Yang et al., 2004), Lantian (Zheng et al., 1992) and Lingtai (Spassov et al., 2003) (MBB: lower part of L8) (Fig. 4). The ChRM directions within a 50-cm-thick Matuyama/Brunhes transition zone switch back and forth several times (Fig. 3). This multiple feature of rapid magnetization changes across the MBB is also discernable in other Chinese loess sections, e.g., Xifeng, Weinan, Lingtai and Luochuan. In Sanmenxia, the upper and lower boundaries of the JNS are encountered at the top of S10 and the boundary of S12/L13, respectively (Fig. 3), in accord with Luochuan and Xifeng (Kukla and An, 1989), but relatively lower than at Weinan and Jingbian (Zhu et al., 1994; Pan et al., 2002; Guo et al., 2002) (JNS: L10-L12) and Baoji (Rutter et al., 1991) (JNS: L10-S12).

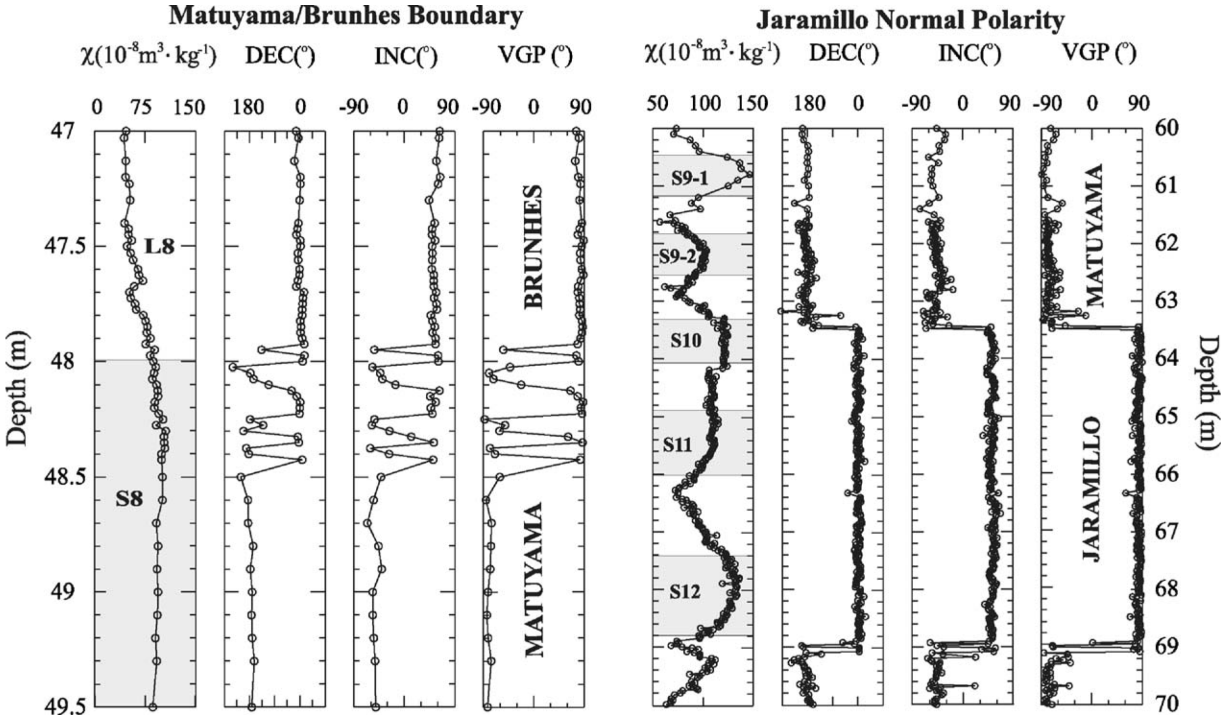

Fig. 3. Stratigraphic variation of mass-specified magnetic susceptibility (χ), ChRM declination, inclination and virtual geomagnetic pole (VGP) latitude across the MBB (left) and JNS (right) at Sanmenxia. The ChRM directions were determined with principal-component analysis using at least four temperature steps for each component. Paleosols are indicated by light shaded regions.

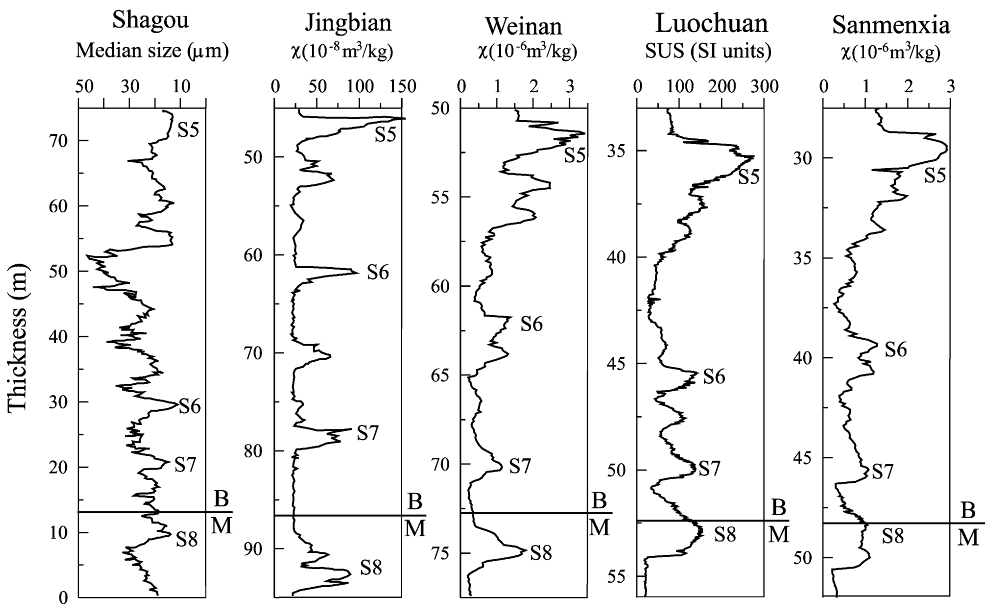

Fig. 4. Median grain size records of the Shagou section (Wu et al., 2005) and magnetic susceptibility records of the Jingbian (Guo et al., 2002; Ding et al., 1999), Weinan (Zhu et al., 2000; Pan et al., 2002), Luochuan (Heller and Liu, 1982; Lu et al., 1999) and Sanmenxia loess sections over the interval of S8-S5. The position of the Matuyama–Brunhes Boundary is indicated in each loess–paleosol sequence.

4. Correlation with $\delta^{18}O$ records

Assuming that the observed MBB and JNS at Sanmenxia represent their true chronostratigraphic positions, we have used a magnetic polarity-based correlation of MS and *S*-ratio records from Sanmenxia to several well-documented MIS records from the North Atlantic (Fig. 5). These integrated high-resolution (high-sedimentation rate) oxygen isotope and paleomagnetic records from ODP Sites 982, 983 and 984, and DSDP Site 607 enable precise correlations. All sites from the North Atlantic show that the MBB and the top of the JNS are encountered within MIS 19 and MIS 27, respectively (Berggren et al., 1995; Channell and Kleiven, 2000; Channell et al., 2002; Channell and Guyodo, 2004). For the base of the JNS, several sites show some inconsistencies, occurring either in the middle of MIS 31 (DSDP 607 and ODP 983) (Berggren et al., 1995; Channell and Kleiven, 2000), the lower part of MIS 31 (ODP 984) (Channell et al., 2002) or the base of MIS 31 (ODP 982) (Channell and Guyodo, 2004).

In contrast to previously reported correlations, we propose that S8, rather than S7 should be correlated with MIS 19 (Zhou and Shackleton, 1999; Heslop et al., 2000). This proposition is justified by the observation that S7 may be correlated to a distinct peak within MIS 18 that is more pronounced in these high-latitude Atlantic cores than in the classic ODP Site 677 (Shackleton et al., 1990). Below the MBB, all peaks and troughs on the MS and *S*-ratio records can be readily correlated to MIS records, and we can confidently correlate S9-1 and S12 to MIS 25 and 31, respectively (Heslop et al., 2000). The MS and *S*-ratio peaks in the middle pedogenic part of L9 are comparable in magnitude to those within S7 and S8. These parameter peaks are also much more prominent compared to those of any previous reports from the hinterland of the CLP, indicating that a distinct pedogenic interval has been developed. Here we tentatively propose that this pedogenic interval in the middle of L9 is correlated to MIS 21 instead of MIS 23 as previously suggested (Heslop et al., 2000; Ding et al., 2002).

5. Discussion

5.1. Large-scale lock-in effects on NRM acquisition in loess?

This proposed correlation does not show obvious discrepancies between the Sanmenxia loess and marine sediments with respect to either the positions of magnetic polarity reversals or paleoclimatic records over the last 1.1 Ma (Fig. 5). This magnetic polarity-based correlation is notably different from previous correlations that assign S7 with MIS 19. Therefore our correlation is not consistent with the commonly adopted lock-in model for Chinese loess (Zhou and Shackleton, 1999; Spassov et al., 2003). The lock-in model predicts that the MBB in Sanmenxia should be shifted downward by about 2.8 m. This would correspond to a time delay of 80 kyr, i.e., the entire duration of the JNS. If this was the case, all polarity reversals, including the entire JNS, would have been correspondingly delayed, and the time delay of remanence acquisition would be larger for the low-accumulation-rate Matuyama loess/paleosols. However, our results do not support any significant downward displacement of the JNS relative to the orbitally tuned marine records (Fig. 5). The lock-in model assumed two separate lock-in zones, one narrow detrital (PDRM/DRM) zone superimposed on a more extensive zone in which a CRM of pedogenic origin is acquired (Spassov et al., 2003). Presently it is well accepted that pedogenesis may result in precipitation of secondary ferrimagnetic phases and a significant enhancement of MS (Zhou et al., 1990). Hence, the lock-in zone should be associated with an increase in MS, which would incorporate the entire L8 in Sanmenxia and most of L8 in other loess sections. However, in Sanmenxia and most other loess sections, MS decreases rather than increases in the lock-in zone, violating the requirements for the model proposed by Spassov et al. (2003). Consequently, the magnetization in Sanmenxia cannot be described by the most recently proposed lock-in model.

Our correlation is also supported by a number of observations that do not support a significant time lag associated with the lock-in model. Firstly, the ChRM in the less pedogenic L8 is thought to be predominantly detrital in origin with a minimal lock-in delay of NRM (Zhu et al., 1994, 1998a), although delayed acquisition of remanence magnetization in the partly pedogenic S7 probably exists. However, detailed paleomagnetic investigations from the Jingbian and Shagou sections in the northern and western margins of the CLP reveal that the MBBs are encountered at the lower part of L8, 5 m and 10 m, respectively below the base of S7 as shown in Fig. 4 (Guo et al., 2002; Pan et al., 2001; Wu et al., 2005). It seems unrealistic that such thick lock-in depths extend from S7 to the lower part of L8 in these weakly pedogenic localities. Secondly, several short-lived geomagnetic excursions in the Brunhes and Matuyama have been identified in different parts of the CLP, implying that the time lag of a remanence acquisition should be less than 10 kyr if a lock-in zone does indeed exist

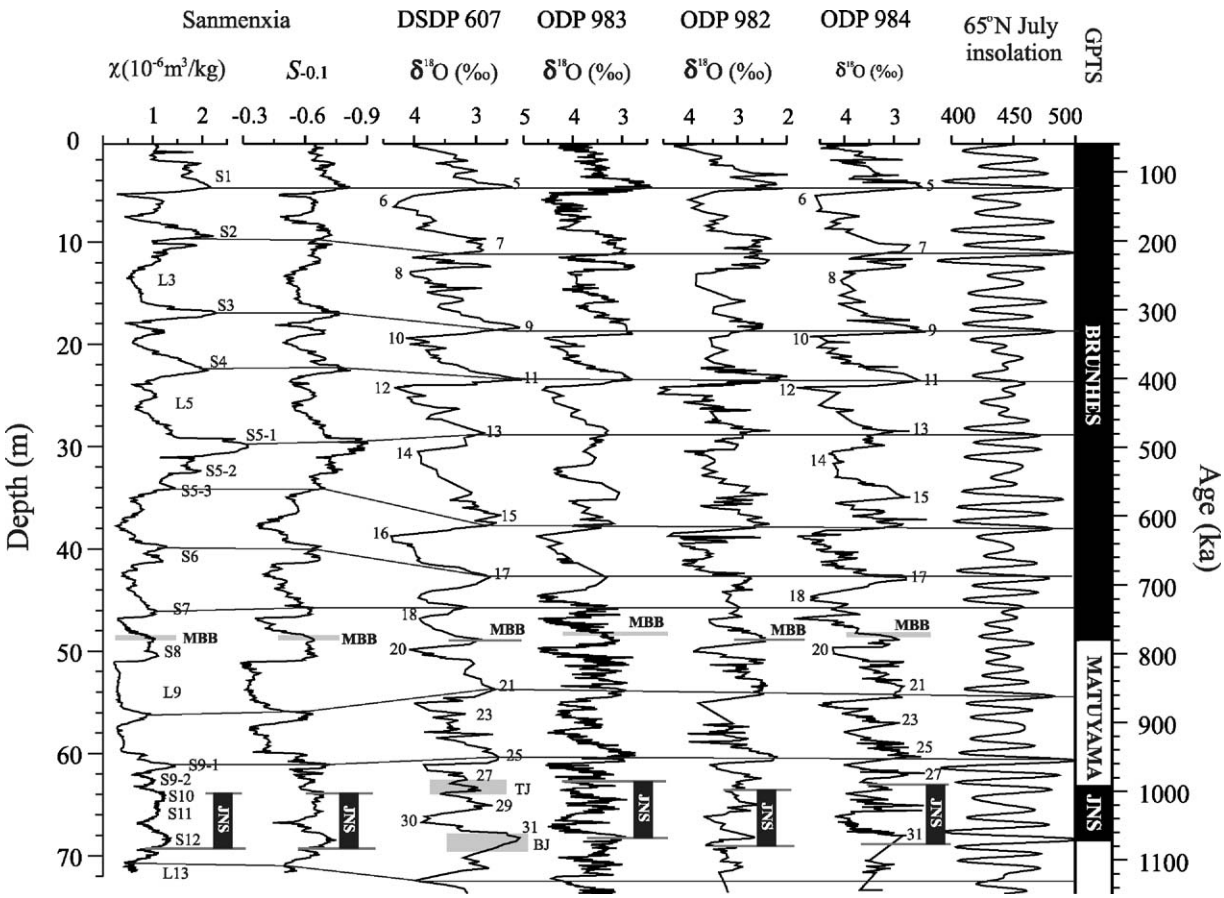

Fig. 5. Proposed correlation of the Sanmenxia magnetic susceptibility and *S*-ratio records to marine isotope records from DSDP Site 607 (Raymo et al., 1989; Ruddiman et al., 1989; Berggren et al., 1995), ODP Sites 983, 982, 984 (Channell and Kleiven, 2000; Channell et al., 2002; Channell and Guyodo, 2004; Raymo et al., 2004) and 65 °N solar insolation curve (Berger and Loutre, 1991). The Sanmenxia paleosols (S*i*) with higher susceptibility are correlated with odd-number oxygen isotope stages as indicated by the thin lines. The position of the MBB in each record is marked with a broad line. Shaded intervals indicate the depth range for the top and bottom of the JNS in DSDP Site 607 core. The geomagnetic polarity time scale (GPTS) is shown on the right of the diagram. Note that the age model for the four Benthic $\delta^{18}O$ records from Atlantic sites is based on correlation to the Shackleton et al. (1990) target record.

(Zhu et al., 1998b, 1999; Pan et al., 2002; Yang et al., 2004). Moreover, a comparison of the loess–paleosol sequence at Lingtai with MIS record over the interval 1.6–2.6 Ma does not reveal any delay of the Olduvai subchron even if a lock-in effect was considered when fine-tuning the ages of the geomagnetic reversal boundaries (Sun and An, 2004). These observations, combined with our results, strongly suggest that extensive lock-in delay is not responsible for the varying positions of the MBB across the CLP.

5.2. The MBB in marine realm

The generally accepted astrochronological age (780 ka) for the MBB was derived using the Ice Volume Model (Imbrie and Imbrie, 1980) as a tuning target for the oxygen isotope data from ODP Site 677 (Shackleton et al., 1990). The Chinese loess timescale is also primarily based on the good correlation of MS or grain size records with the MIS record from ODP 677 (Heller and Evans, 1995; Heslop et al., 2000; Ding et al., 2002). However, ODP Site 677 does not have a magnetic polarity stratigraphy and the age of reversals including the MBB was deduced by correlation to DSDP Hole 522A and DSDP 607 (Shackleton et al., 1990). Therefore the interpretation of MBB inconsistencies between Chinese loess and marine record from ODP 677 is not an unambiguous exercise.

The well-documented paleomagnetic and isotopic records from the North Atlantic provide an opportunity to match MS signals from Chinese loess to MIS records at higher resolution than has been possible previously. At

ODP Sites 983 and 984 (Iceland Basin), the mean sedimentation rates in the 700–900 ka interval are 13 cm/kyr and 12 cm/kyr, respectively, which are much higher than ODP Site 677 (4 cm/kyr over the last 1 Ma) (Shackleton et al., 1990; Channell and Kleiven, 2000; Channell et al., 2004). Both the MBBs at ODP Sites 983 and 984 occur at the young end of MIS 19, with midpoint ages at 772.5 ka and 773.5 ka, respectively (Fig. 5). These ages are younger than the generally adopted astrochronological estimate (780 ka) even if these high-resolution paleomagnetic records are also apparently affected by PDRM smoothing and downward displacement of the MBB. Since the effective lock-in depth for remanence acquisition may have greater temporal significance at sites with a lower sedimentation rate (from which the astrochronological estimates have been derived) than at sites with a higher sedimentation rate such as ODP Sites 983 and 984, the younger ages of the MBB at Sites 983 and 984 could be attributed to a finite lock-in depth for magnetic remanence acquisition (Channell and Kleiven, 2000; Channell and Guyodo, 2004).

We should also note that marine sediments are usually extensively bioturbated to depths ranging from a few centimeters to tens of centimeters below the sediment/water interface. The bioturbation will therefore simply cause a delay in PDRM acquisition (Irving and Major, 1964; Kent, 1973; Løvlie, 1974; Roberts and Winklhofer, 2004). By contrast, eolian deposits like loess are not in general affected by bioturbation or disturbances by water, and post-depositional alterations of loess–paleosol sequence are likely to be minimal. Although weathering/pedogenesis after deposition may promote oxidation and create partly oxidized magnetic phases that may carry significant magnetic directions, this secondary component presumably representing low-temperature oxidation of detrital magnetite together with a viscous remanent magnetization may be eliminated by thermal treatment at peak temperatures between 250 °C and 300 °C (Fig. 2). After removal of the heavy overprints, the well-defined ChRM directions are carried predominantly by magnetite, which is supposed to be detrital in origin. The recording consistency of the MBB in the Sanmenxia loess and high-sedimentation-rate marine records further suggests that the effective time lag in remanence acquisition in loess is probably comparable to that in subaqueous marine sediments, although they are formed in different sedimentary environments and from different provenance.

Moreover, the average accumulation rate within the Brunhes chron at Sanmenxia amounts to 48 m/780 kyr, implying that the 50-cm-thick Matuyama/Brunhes polarity transition has an apparent duration of 7.7 kyr. This estimate is comparable to that at ODP Sites 983 (5 kyr) and 984 (7 kyr) (Channell and Kleiven, 2000; Channell et al., 2004). Therefore we propose that the MBBs appearing as multiple polarity flips in Chinese loess essentially reflect true features of the (local) geomagnetic field and cannot be attributed to artifacts of the recording process. Inconsistencies in stratigraphic positions and directional records associated with the MBB encountered in different loess sections across the CLP may be attributed to local variations in accumulation rates and pedogenesis causing varying degrees of directional smoothing, loss of resolution due to non-continuous (discrete) sampling strategies and dissimilar demagnetization/analytical techniques in different studies.

5.3. Time transgressive loess/paleosol boundaries across the CLP

The inconsistent occurrence of the MBB at the top of S8, lower part or the bottom of L8 on the CLP may be attributed to regional/local variability in loess/paleosol development across the CLP responding to oscillations of the winter and summer monsoons. Undoubtedly, the winter (summer) monsoon is much more vigorous (weaker) at the northwestern margin of the CLP than at the southeastern margin and vice versa. Field observations indicate that the thickness of a certain loess layer in the northern part of the CLP is much greater than that in the southern part of the CLP (Liu, 1985). Consequently, during the culminating development of S8 at the southeastern extremity of the CLP, probably the base of L8 commenced accumulating at the northern part of the CLP. It turns out that the MBB occurs during the terminal stage of S8 formation in the southeastern margin of the CLP, whereas it falls in the bottom or lower part of L8 in the hinterland and NW of the CLP. We also believe that corresponding peaks/troughs of MS records in a given soil/loess unit throughout the CLP roughly mark the same age. But this does not necessarily imply that the age of the top of S8 is exactly consistent throughout the CLP. The prominent MS peak in L9 in Sanmenxia also suggests that loess deposition and soil formation are competing processes, and that the magnetic signature in loess/soil is the result of these competing processes. Therefore we probably need to reconsider the synchronization of the base/top for certain loess/paleosol units across the CLP. Under such circumstances, we can also tentatively attribute the MBB discrepancy between marine sediments and most loess records from the hinterland and north part the CLP to a more rapid response of the CLP compared with the high-heat-capacity oceans to initial cooling and glacial expansion across the interglacial–glacial (stage 19-18) transition, although presently very little is known about the interhemispheric phase relationships between the continents and oceans to climatic change. Our views that loess/paleosol boundaries across the CLP are not strict chronostratigraphic boundaries and there could be a land–ocean climatic recording phase lag deserve further multi-discipline investigations.

6. Conclusions

Relying on high-resolution paleomagnetic constraints of loess/paleosol sequences from the Sanmenxia section at the SE extremity of the CLP, we present an alternative paleoclimate comparison between Chinese loess and MIS record over the last 1.1 Ma. This revised correlation shows that the transgressive nature of climate change and hence loess/paleosol boundaries across the CLP are the primary factors responsible for the discordant positions of polarity reversal boundaries. Consequently, previously proposed large-scale lock-in effects of loess magnetization acquisition associated with pedogenesis have probably been overestimated. Based on our correlations, S8 in Sanmenxia should correspond to the interglacial MIS 19 while not to the hitherto supposed MIS 21, and a distinct second-order peak within MIS 18 may be correlated to S7. Our results also suggest that the "upper sand layer"-L9 does not necessarily represent a complete cold extreme, since it has evidently been interrupted by a distinct pedogenic interval that should be correlated to MIS 21 instead of MIS 23.

Acknowledgements

We thank two anonymous reviewers for their constructive criticisms that helped to improve the clarity of this manuscript. Profs. R.X. Zhu and F.C. Jiang are greatly appreciated for their valuable suggestions. We are also grateful to Prof. M.E. Raymo for providing benthic $\delta^{18}O$ data set. This study is funded by the NSFC (40202018), China Geological Survey (200413000035), the Key Laboratory of Crust Deformation and Processes, CAGS (KL05-13) and the Ministry of Personnel of China.

References

An, Z.S., 2000. The history and variability of the East Asian paleomonsoon climate. Q. Sci. Rev. 19, 171–187.

Berger, A., Loutre, M.F., 1991. Insolation values for the climate of the last 10 million years. Q. Sci. Rev. 10, 297–317.

Berggren, W.A., Hilgen, F.J., Langereis, C.G., Kent, D.V., Obradovich, J.D., Isabella Raffi, Raymo, M.E., Shackleton, N.J., 1995. Late Neocene chronology: new perspective in high-resolution stratigraphy. Geol. Soc. Am. Bull. 107, 1272–1287.

Channell, J.E.T., Kleiven, H.F., 2000. Geomagnetic palaeointensities and astrochronological ages for the Matuyama–Brunhes boundary and the boundaries of the Jaramillo Subchron: palaeomagnetic and oxygen isotope records from ODP Site 983. Phil. Trans. R. Soc. Lond. A 358, 1027–1047.

Channell, J.E.T., Mazaud, A., Sullivan, P., Turner, S., Raymo, M.E., 2002. Geomagnetic excursions and paleointensities in the 0.9–2.15 Ma interval of the Matuyama Chron at ODP Site 983 and 984 (Iceland Basin). J. Geophys. Res. 107, doi:10.1029/2001JB000491.

Channell, J.E.T., Guyodo, Y., 2004. The Matuyama Chronozone at ODP Site 982 (Rockall Bank): evidence for decimeter-scale magnetization lock-in depths. In: Channell, J.E.T., Kent, D.V., Lowrie, W., Meert, J. (Eds.), Timescales of the Paleomagnetic Field. Am. Geophys. Union, Geophys. Monogr. No. 145, 328 pp.

Channell, J.E.T., Curtis, J.H., Flower, B.P., 2004. The Matuyama–Brunhes boundary interval (500–900 ka) in North Atlantic drift sediments. Geophys. J. Int. 158, 489–505.

Ding, Z.L., Yu, Z., Rutter, N.W., Liu, T.S., 1994. Towards an orbital time scale for Chinese loess deposits. Q. Sci. Rev. 13, 39–70.

Ding, Z.L., Sun, J.M., Liu, T.S., 1999. Stepwise advance of the Mu Us desert since late Pliocene: evidence from a red clay–loess record. Chin. Sci. Bull. 44, 1211–1214.

Ding, Z.L., Derbyshire, E., Yang, S.L., Yu, W., Xiong, S.F., Liu, T.S., 2002. Stacked 2.6 Ma grain size record from the Chinese loess based on five sections and correlation with the deep-sea $\delta^{18}O$ record. Paleoceanography 17, doi:10.1029/2001PA000725.

Evans, M.E., Heller, F., 2001. Magnetism of loess/paleosol sequence: recent developments. Earth Sci. Rev. 54, 129–144.

Guo, B., Zhu, R.X., Florindo, F., Ding, Z.L., Sun, J.M., 2002. A short reverse polarity interval within the Jaramillo subchron: evidence from the Jingbian section, northern Chinese Loess Plateau. J. Geophys. Res. 107 (B6), doi:10.1029/2001JB000706.

Heller, F., Liu, T.S., 1982. Magnetostratigraphic dating of loess deposits in China. Nature 300, 431–433.

Heller, F., Evans, M.E., 1995. Loess magnetism. Rev. Geophys. 33, 211–240.

Heslop, D., Langereis, C.G., Dekkers, M.J., 2000. A new astronomical timescale for the loess deposits of Northern China. Earth Planet. Sci. Lett. 184, 125–139.

Hus, J.J., Han, J.M., 1992. The contribution of loess magnetism in China to the retrieval of past global changes-some problems. Phys. Earth Planet. Int. 70, 154–168.

Imbrie, J., Imbrie, J.Z., 1980. Modeling the climatic response to orbital variations. Science 207, 943–953.

Irving, E., Major, A., 1964. Post-depositional detrital remanent magnetization in a synthetic sediment. Sedimentology 3, 135–143.

Kent, D.V., 1973. Post depositional remanent magnetization in deep sea sediments. Nature 246, 32–34.

Kukla, G., Heller, F., Liu, X.M., Xu, T.C., Liu, T.S., An, Z.S., 1988. Pleistocene climates in China dated by magnetic susceptibility. Geology 16, 811–814.

Kukla, G., An, Z.S., 1989. Loess stratigraphy in central China. Palaeogeogr. Paleoclimatol. Paleoecol. 72, 203–225.

Liu, T.S., 1985. Loess and the environment. China Ocean Press, Beijing, 215 pp.

Liu, T.S., Ding, Z.L., 1998. Chinese loess and the palaeomonsoon. Ann. Rev. Earth Planet. Sci. 26, 111–145.

Liu, X.M., Liu, T.S., Xu, T.C., Liu, C., Cheng, M.Y., 1987. A preliminary study on magnetostratigraphy of a loess profile in Xifeng area, Gansu Province. In: Liu, T.S. (Ed.), Aspects of Loess Research. China Ocean Press, Beijing, pp. 64–174.

Løvlie, R., 1974. Post-depositional remanent magnetization in redeposited deep-sea sediment. Earth Planet. Sci. Lett. 21, 315–320.

Lu, H.Y., Liu, X.D., Zhang, F.Q., An, Z.S., Dodson, J., 1999. Astronomical calibration of loess–paleosol deposits at Luochuan, Central Chinese Loess Plateau. Palaeogeogr. Paleoclimatol. Paleoecol. 154, 237–246.

Pan, B.T., Wu, G.J., Wang, Y.X., Liu, Z.G., Guan, Q.Y., 2001. Age and genesis of Shagou River Terraces in Eastern Qilian Mountains. Chin. Sci. Bull. 46, 509–513.

Pan, Y.X., Zhu, R.X., Liu, Q.S., Guo, B., Yue, L.P., Wu, H.N., 2002. Geomagnetic episodes of the last 1.2 Myr recorded in Chinese loess. Geophys. Res. Lett. 29 (8), doi:10.1029/2001GL014024.

Raymo, M.E., Ruddiman, W.F., Backman, J., Clement, B.M., Martinson, D.G., 1989. Late Pliocene variation in Northern Hemisphere ice sheets and North Atlantic deep circulation. Paleoceanography 4, 413–446.

Raymo, M.E., Oppo, D.W., Flower, B.P., Hodell, D.A., McManus, J.F., Venz, K.A., Kleiven, K.F., McIntyre, K., 2004. Stability of North Atlantic water masses in face of pronounced climate variability during the Pleistocene. Paleoceanography 19, doi:10.1029/2003PA000921.

Roberts, A.P., Winklhofer, M., 2004. Why are geomagnetic excursions not always recorded in sediments? Constraints from post-depositional remanent magnetization lock-in model. Earth Planet. Sci. Lett. 227, 345–359.

Ruddiman, W.F., Raymo, M.E., Martinson, D.G., Clement, B.M., Backman, J., 1989. Pleistocene evolution: northern hemisphere ice sheets and North Atlantic Ocean. Paleoceanography 4, 353–412.

Rutter, N., Ding, Z.L., Evans, M.E., Liu, T.S., 1991. Baoji-type pedostratigraphic section, loess plateau, north-central China. Q. Sci. Rev. 10, 1–22.

Shackleton, N.J., Berger, A., Peltier, W.R., 1990. An alternative astronomical calibration of the Lower Pleistocene timescale based on ODP Site 677. Trans. R. Soc. Edinburgh Earth Sci. 81, 251–261.

Spassov, S., Heller, F., Evans, M.E., Yue, L.P., von Dobeneck, T., 2003. A lock-in model for the complex Matuyama/Brunhes boundary record of the loess/paleosol sequence at Lingtai (Central Chinese Loess Plateau). Geophys. J. Int. 155, 350–366.

Sun, D.H., Shaw, J., An, Z.S., Cheng, M., Yue, L.P., 1998. Magnetostratigraphy and paleoclimatic interpretation of a continuous 7.2 Ma Late Cenozoic eolian sediments from the Chinese Loess Plateau. Geophys. Res. Lett. 25, 85–88.

Sun, Y.B., An, Z.S., 2004. An improved comparison of Chinese loess with deep-sea δ^{18}O record over the interval 1.6–2.6 Ma. Geophys. Res. Lett. 31, doi:10.1029/2004GL019716.

Tauxe, L., Herbert, T., Shackleton, N.J., Kok, Y.S., 1996. Astronomical calibration of the Matuyama–Brunhes boundary: consequences from magnetic remanence acquisition in marine carbonates and the Asian loess sequences. Earth Planet. Sci. Lett. 140, 133–146.

Wang, X., Løvlie, R., Yang, Z., Pei, J., Zhao, Z., Sun, Z., 2005. Remagnetization of quaternary eolian deposits: a case study from SE Chinese Loess Plateau. Geochem. Geophys. Geosyst. 6, Q06H18, doi:10.1029/2004GC000901.

Wu, G.J., Pan, B.T., Guan, Q.Y., Xia, D.S., 2005. Terminations and their correlation with solar insolation in the Northern Hemisphere: a record from a loess section in North China. Palaeogeogr. Paleoclimatol. Paleoecol. 216, 267–277.

Xiong, S.F., Ding, Z.L., Liu, T.S., 2001. Climatic implications of loess deposits from the Beijing region. J. Q. Sci. 16, 575–582.

Yang, T.S., Hyodo, M., Yang, Z., Fu, J., 2004. Evidence for the Kamikatsura and Santa Rosa excursions recorded in eolian deposits from the southern Chinese Loess Plateau. J. Geophys. Res. 109, B12105, doi:10.1029/2004JB002966.

Zheng, H.B., An, Z.S., Shaw, J., 1992. New contributions to Chinese Plio-Pleistocene magnetostratigraphy. Phys. Earth Planet. Int. 70, 146–153.

Zhou, L.P., Oldfield, F., Wintle, A.G., Robinson, S.G., Wang, J.T., 1990. Partly pedogenic origin of magnetic variations in Chinese loess. Nature 346, 737–739.

Zhou, L.P., Shackleton, N.J., 1999. Misleading position of geomagnetic reversal boundaries in Eurasian loess and implications for correlation between continental and marine sedimentary sequences. Earth Planet. Sci. Lett. 168, 117–130.

Zhu, R.X., Laj, C., Mazaud, A., 1994. The Matuyama–Brunhes and Upper Jaramillo transitions recorded in a loess section at Weinan, north-central China. Earth Planet. Sci. Lett. 125, 143–158.

Zhu, R.X., Pan, Y.X., Guo, B., Liu, Q.S., 1998a. A recording phase lag between ocean and continent climate change: constrained by the Matuyama/Brunhes polarity boundary. Chin. Sci. Bull. 49, 1593–1598.

Zhu, R.X., Coe, R.S., Guo, B., Anderson, R., Zhao, X.X., 1998b. Inconsistent palaeomagnetic recording of the Blake event in Chinese loess related to sedimentary environment. Geophys. J. Int. 134, 867–875.

Zhu, R.X., Pan, Y.X., Liu, Q.S., 1999. Geomagnetic excursions recorded in Chinese loess in the last 70 000 yr. Geophys. Res. Lett. 26, 505–508.

Zhu, R.X., Guo, B., Ding, Z.L., Guo, Z.T., Kazansky, A., Matasova, G., 2000. Gauss-Matuyama polarity transition obtained from a loess section at Weinan, North-central China. Chin. J. Geophys. 43, 654–671.