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Magnetism of Chinese loess deposits

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Summary. The age of Chinese loess deposits has long been disputed. Biostratigraphical and earlier magnetostratigraphical investigations placed the entire loess formation within the Pleistocene and ascertained a maximum loess age of about 1.2 Myr. A new collection of nearly 500 samples from a natural outcrop and a borehole section near Lochuan (lat. 35.8 °N, long. 109.2 °E; Shaanxi province) has been dated by magnetic stratigraphy. Thermal cleaning of the natural remanent magnetization (NRM) removes a strong secondary component of viscous origin along the present geomagnetic field which resides largely in magnetite. The characteristic NRM component is due to haematite which is thought to be of chemical origin. Rhythmical intensity variations of NRM and initial susceptibility depend on the loess lithology and may reflect climatic changes during loess deposition.

The palaeomagnetic results are consistent between the two sections and yield a clearly defined magnetic polarity zonation. The Brunhes–Matuyama boundary and the Jaramillo subchron have been positively identified in both outcrops at exactly the same stratigraphic level. The Olduvai subchron has been found in the borehole section which records the entire loess sequence. Most probably the formation of Chinese loess began shortly after the Matuyama–Gauss polarity transition. Therefore a late Pliocene age of about 2.4 Myr is assigned to the oldest loess sediments measured.

Introduction

The widespread occurrence of loess – more than 6 per cent of the territory of China is covered by 'huangtu' (yellow earth) – has attracted the attention of Chinese earth scientists for a long time. As early as 2000 years ago the concept of an aeolian origin of loess has been put forward in China (Wang & Zhang 1980). The mineral grains making up these silty sediments originated from subglacial grinding processes (Smalley 1975) or from disintegration of source rocks due to diurnal temperature changes followed by redistribution and deposition by the wind.

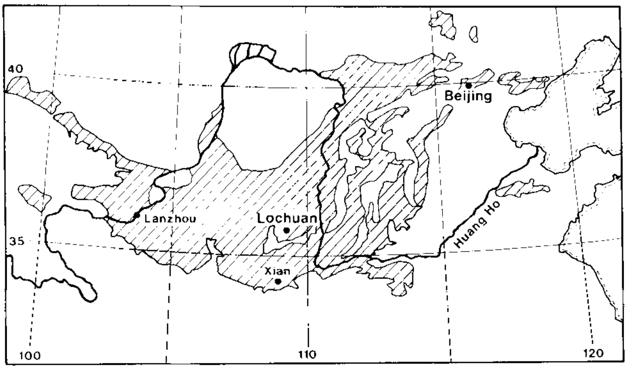

Figure 1. Geological sketch map of the loess plateau with the sampling sites near Lochuan (redrawn from Wang & Zhang 1980).

The Chinese loess sediments were deposited in the arid and semi-arid regions of an E–W-trending girdle between 33° and 47° northern latitude and between 127° and 75° eastern longitude. The central part of this great belt is formed by the well-known loess plateau which occupies a coherent area of more than 300 000 km^2 (Fig. 1). Along the middle Huang Ho (Yellow River) valley loess thicknesses are generally 80–120 m and in the county of Lochuan (Shaanxi province) the pre-loessic topography allows for a maximum thickness of 175 m (Liu & Chang 1964).

One of the major geological problems is the dating of the Chinese loess. Mammalian and invertebrate fossils give only rough absolute age estimates which associate the oldest sediments in the formation with the early to late Pleistocene. Previously available thermoluminescence ages have been of questionable value. Magnetostratigraphic investigations which relate the polarity of loess natural remanent magnetization (NRM) to the well-known geomagnetic polarity changes during the last few million years (Mankinen & Dalrymple 1979), have been initiated by several groups of Chinese scientists (Li, An & Wang 1974; An, Wang & Li 1977; Wang *et al.* 1980). They believe to have identified the Brunhes–Matuyama (B/M) boundary as well as the Jaramillo subchron in several natural outcrops or core sections from boreholes and conclude that the maximum age of loess formation in China ranges between 1.0 and 1.2 Myr BP.

In the above studies the characteristic remanent magnetization (ChRM) directions were established by alternating field (AF) demagnetization. This technique resulted in an asymmetric ChRM inclination distribution in which the reversed inclinations are considerably shallower and much more scattered than the normal inclinations. The data seem to be strongly influenced by secondary magnetization components associated with high coercivity ferromagnetic minerals which could not be completely demagnetized by the AF technique. As Davies *et al.* (1977) pointed out, this type of overprint often affects the NRM properties of Quaternary sediments. A certain number of short-term excursions of either polarity – usually defined by only one sample – were observed at different stratigraphic positions in the loess. Often these cannot be correlated between sections, and their significance as reliable palaeomagnetically datable marker horizons is dubious.

Palaeomagnetic investigations of Central European loess units (Bucha & Horacek 1973; Kukla 1975; Brunnacker 1977; Fink & Kukla 1977; Tucholka 1977) suggest that loess deposition in Europe started earlier, at least 1.9 Myr ago. The quality of the magnetic data in these studies requires considerable improvement of NRM cleaning. Often

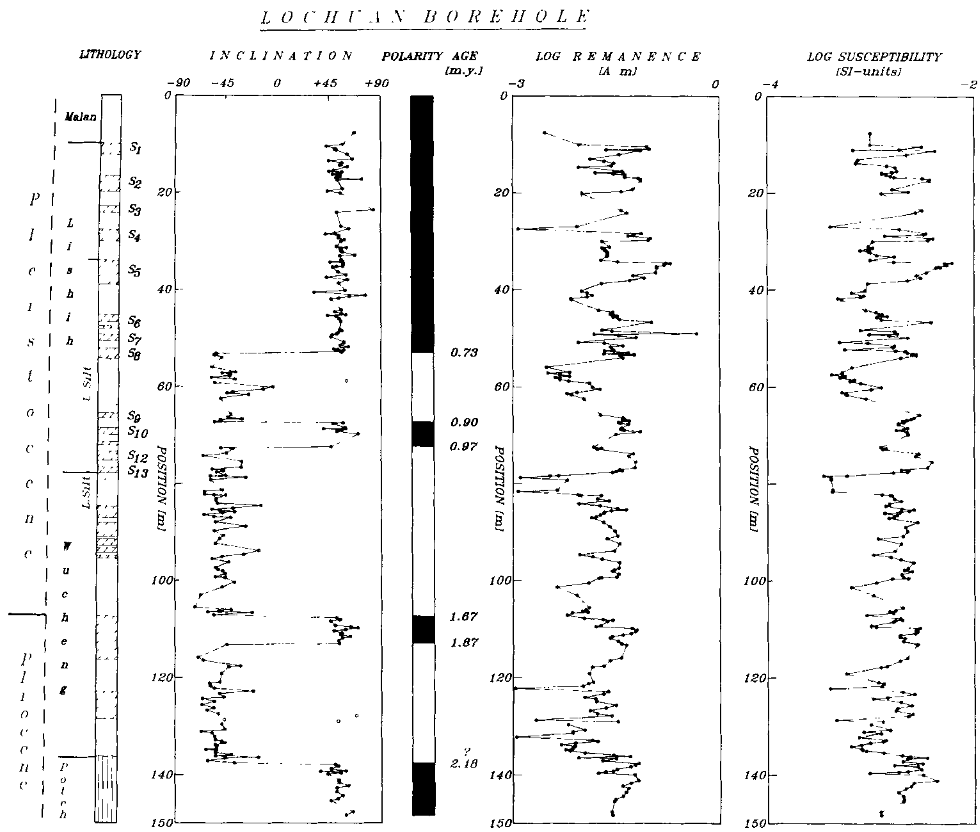

Figure 2. Stratigraphic dependence of the thermally cleaned ChRM inclinations, and the NRM intensity and initial susceptibility of the unheated samples from the Lochuan borehole. The polarity column with absolute time marks is according to Mankinen & Dalrymple (1979). Key to lithology: loess layers are white, silt layers are stippled and the numbering of hatched soil layers is according to Liu *et al.* (1980).

'abnormal' directions having deviating declinations, but still normal inclinations define a polarity event. However, the identifications of the Jaramillo and Olduvai subchrons seem to be reliable. The onset of European loess sedimentation was roughly contemporary with a change towards colder climate in the northern hemisphere. Evidence for this happening during the early Matuyama polarity chron (2.4 Myr) comes from the oxygen isotope record of palaeomagnetically dated deep sea sediments (Shackleton & Opdyke 1977).

The different ages for the commencement of loess formation (*c.* 1.2 or 1.9 Myr) and major climatic changes (2.4 Myr) need further investigation. In a recent paper Heller & Liu (1982) have summarized magnetostratigraphical data obtained from a borehole section near Lochuan (Shaanxi province) *c.* 800 km SW of Beijing (Fig. 1). These new data yielded an approximate age of *c.* 2.4 Myr for the onset of Chinese loess deposition, in good agreement with the deep sea record and to a lesser extent with the European loess studies.

Sedimentology and stratigraphy

Chinese loess deposits typically consist of alternations of silty loess layers and clayey loess or palaeosoil beds (Fig. 2). Depending on the extent of weathering pedogenesis the loess horizons, which are devoid of stratification or lamination, vary in colour from greyish-yellow to light red. Due to advanced weathering their calcium carbonate content becomes reduced and the clay content increases (Liu *et al.* 1980). The palaeosoils of parabraunerde and braunerde type are greyish-brown to dark red. They are strongly depleted in calcium carbonate and enriched with clay minerals. *Cathaicae* (gastropod) assemblages in the loess layers indicate an environment of dry and cold steppe during loess deposition, whereas *Metodontia* assemblages observed in the palaeosoils indicate a relatively warm and humid climate during soil formation (Chen *et al.* 1979). The environmental changes are reflected similarly by variations in crystallinity and montmorillonite content of illite, the main clay mineral in loess (Han *et al.* 1979), as well as by other geochemical parameters such as FeO/Fe_2O_3 ratios (Wen *et al.* 1979). Thus it has been concluded that the alternation of loess and palaeosoil horizons was largely caused by climatic variations. The loess layers represent climatic glacial stages and the soils formed during interglacial stages (Liu *et al.* 1980).

Heavy minerals make up 4–7 per cent by volume of the loess mineral content. Iron oxides and hydroxides constitute 20–40 per cent of this fraction. Magnetite predominates with 10–20 per cent, limonite follows with 7–10 per cent, haematite contributes 1–8 per cent and ilmenite 1–5 per cent. X-ray diffractograms of loesses and soils show peaks which also correspond to magnetite, haematite and ilmenite, but goethite (limonite) has not been positively identified by this method. In polished sections (Fig. 3) the magnetite is seen to be oxidized to haematite to a variable extent. The oxidation is more pronounced near grain rims than in the grain interiors. This suggests that the oxidation occurred at low temperatures during or after sediment deposition. Haematite also forms from a goethite precursor (Schwertmann 1966), the remainders of which could be identified in a few cases from the dull grey to bluish colour. Some dark minerals, which might be hornblende or mica, contain very fine grained exsolutions of a highly reflecting ore mineral which is probably again haematite.

On the basis of lithological characteristics, unconformable contacts and erosional surfaces, the loess has been subdivided stratigraphically into three main members all of which have been assigned a Pleistocene age (Teilhard de Chardin & Young 1930; Kes 1958; Yen 1960; Liu & Chang 1964). Following the sequence from top to bottom, the youngest sediments below the Holocene loess belong to the Malan loess (Fig. 2) with mammalian fossils typical of late Pleistocene age. The underlying Lishi member, which is usually divided by an erosional surface into Upper and Lower Lishi, contains elements of Zhoukoudian faunas (equivalent to the European Oldenburgian). According to fission track dating (Guo *et al.* 1980) this formation has an age of 460 000 yr which places the member in the Middle Pleistocene. The oldest loess unit, the Wucheng loess, contains the Villefranchian Nihowan fauna considered to be early Pleistocene. The pre-loessic sediments are often constituted by the Poteh red clays which have recently been dated palaeontologically as late Pliocene. The thickness of the individual loess members varies considerably throughout the loess plateau depending on position with respect to the source area in the north, on pre-loessic topography and on local erosion during loess formation.

The borehole section near Lochuan comprises a 136 m thick sequence of loess and soil layers (Fig. 2). The Malan, Upper Lishi, Lower Lishi and Wucheng have thicknesses of 9.8, 25.2, 43.8 and 58.1 m, respectively. The Upper Pliocene red clay at the bottom is 9.8 m thick. The same unit thicknesses have been measured in a natural outcrop nearby (Fig. 4), but here the Lower Lishi member contains higher soil portions. More than 250 samples were obtained from the core with an average sampling interval of 60 cm. The samples were oriented relative to the vertical but not azimuthally. Another 238 samples were

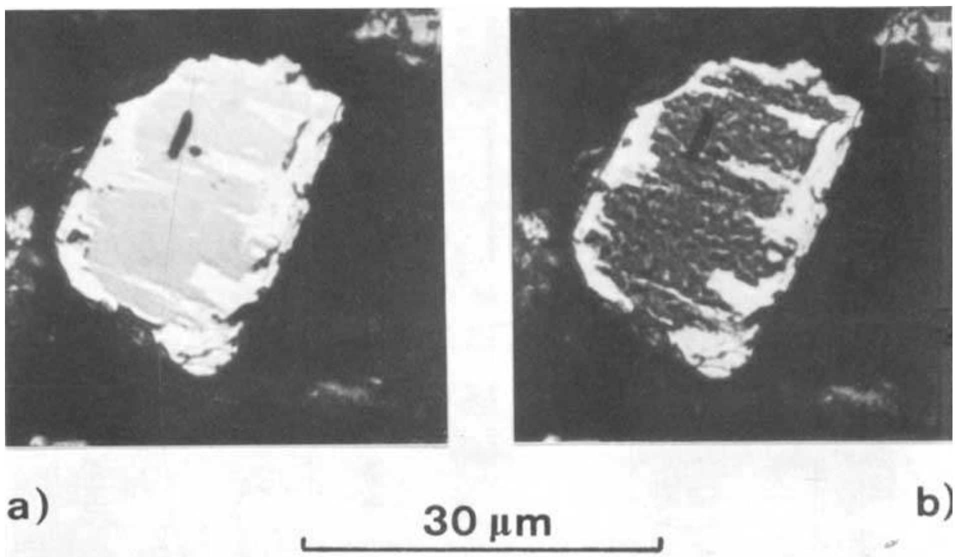

Figure 3. Polished sections of ferromagnetic loess extract. (a) The grey magnetite grain is oxidized to white haematite mainly along the grain rims. (b) Magnetic colloid is attracted by the strongly magnetized magnetite portions (now dark) of the grain, whereas the weakly magnetized haematite (still white) is not covered by the colloid.

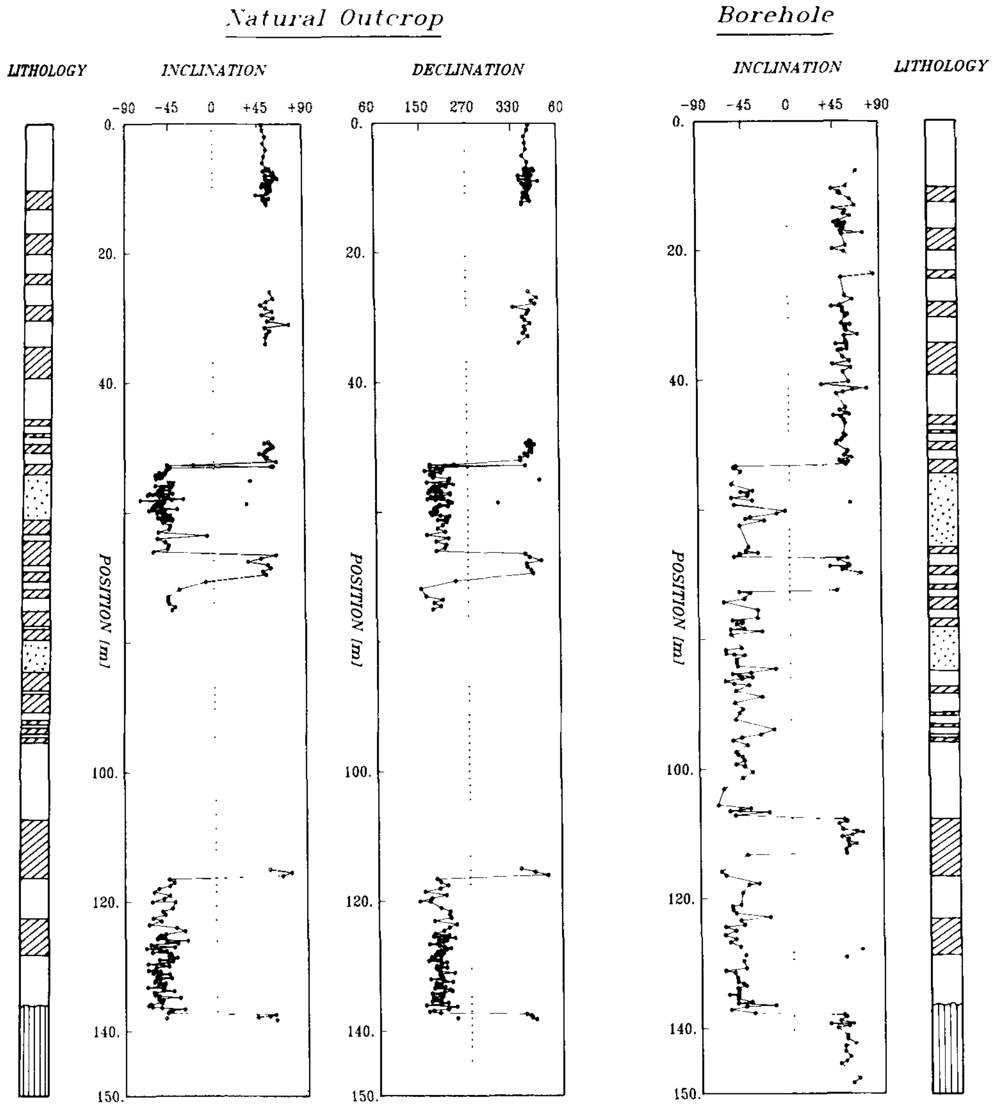

Figure 4. A comparison of the stratigraphic variations of declination, inclination and lithology of the natural outcrop with the borehole inclination and lithology.

collected from the natural outcrop, the sampling being restricted to sections where reversal boundaries had been found in the long core or where polarity events of short duration presumably should have occurred. For this purpose an average sampling interval of 20–25 cm corresponding to *c.* 3000–4000 yr continuous deposition was chosen.

Remanent magnetization and initial susceptibility

The ferromagnetic mineralogy of the loess sediments has been determined magnetically by analysing the coercivity and unblocking temperature spectra (Heller 1978) of isothermal remanent magnetization (IRM). The IRM behaviour of both loess and soil deposits is not substantially different, but the intensities of saturation IRM acquired in a field of 1 T

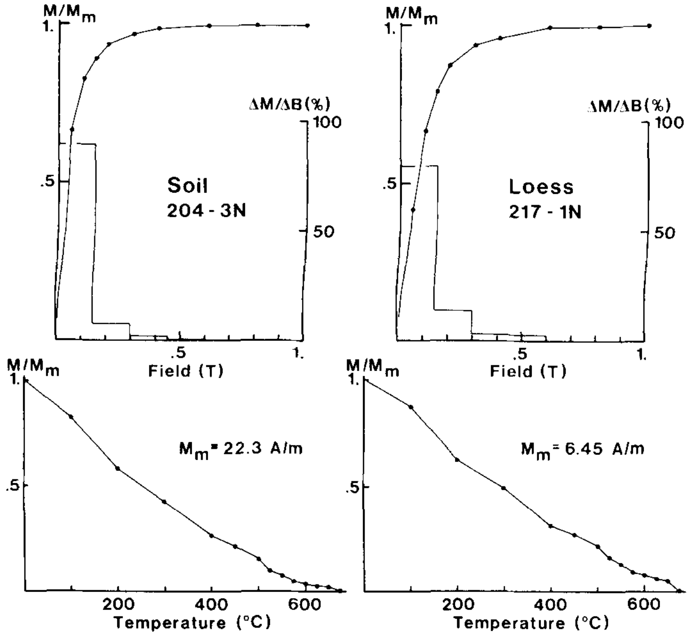

Figure 5. Acquisition and progressive thermal demagnetization of isothermal remanent magnetization (IRM) in loess and palaeosoil samples. The saturation IRM of the palaeosoils is typically higher than that of the loess by a factor of 3–4.

are higher by a factor 3–4 in the plaeosoils (Fig. 5). The major proportion of IRM is always carried by a low coercivity mineral fraction which is attributed to magnetite; this is supported by reflected light microscopic examination of polished sections (Fig. 3). The kinks in the thermal demagnetization curves of IRM between 525 and 575°C also reflect the presence of magnetite. There are always small, but distinct contributions to the IRM acquisition curve from high coercivity material [$(B_0)_{cr} > 0.3$ T] and thermal demagnetization of saturation IRM has maximum unblocking temperatures between 650 and 675°C. This is interpreted as evidence for haematite which also has been identified positively under the microscope.

The intensities of natural remanent magnetization (NRM) and initial susceptibility are generally higher by a factor of 3–4 in the soils than in the loess horizons. This was also noted in the earlier work of Li *et al.* (1974) and An *et al.* (1977). The NRM intensities have geometric mean values of 4.19×10^{-2} A m^{-1} for 178 soil samples, and 1.21×10^{-2} A m^{-1} for 227 loess samples. The low field susceptibility of the soil horizons (mean: 2.31×10^{-3} SI units) is also about three times that of the loess layers (mean: 8.94×10^{-4} SI units). The soils develop from loess by a process of weathering and pedogenetic alteration whereby carbonate content and porosity are largely reduced. We attribute the higher soil intensities mainly to relative enrichment of magnetite. The eventual new formation of other ferromagnetic minerals such as goethite and haematite will not influence the magnetic intensity parameters appreciably because their spontaneous magnetizations are very low compared to that of magnetite.

Detailed progressive demagnetization studies show that the NRM always has been overprinted by a strong secondary component of normal polarity (Fig. 6). Alternating field

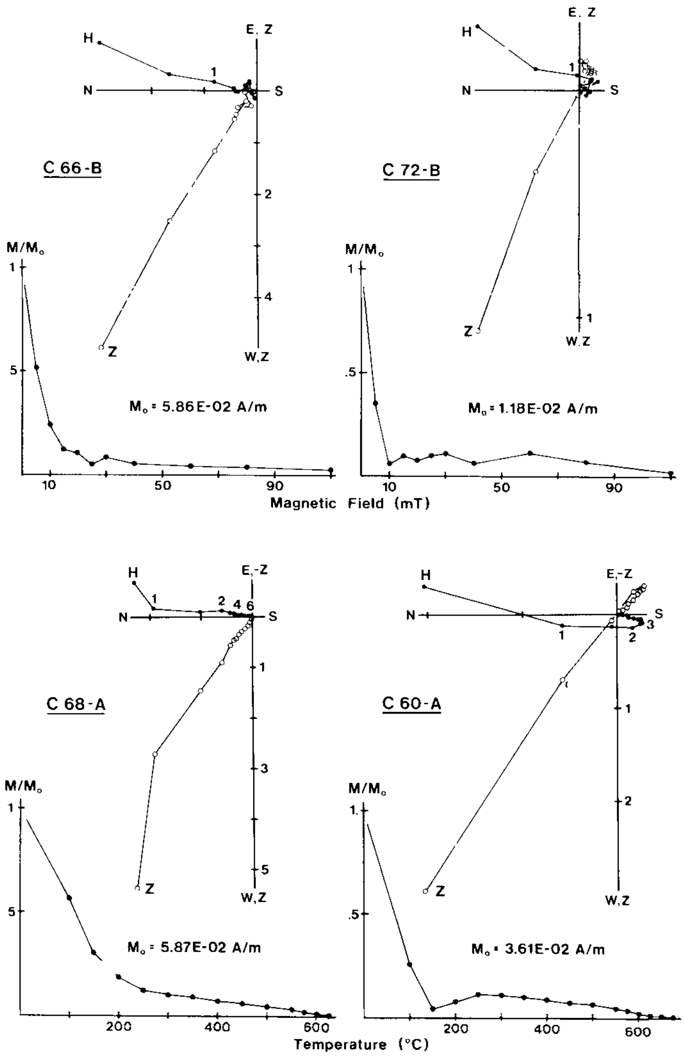

Figure 6. Orthogonal projections of NRM vector paths and NRM intensity behaviour of two palaeosoil (C66-B, C68-A) and two loess samples (C72-B, C60-A) during AF and thermal demagnetization. Numbers along horizontal NRM component (*H*) in vector plots refer to alternating fields/10 (mT) and temperatures/100 (°C) respectively.

(AF) demagnetization up to only 10 mT and thermal cleaning up to 200°C removes about 90 per cent of the initial NRM intensity; most of the secondary magnetization apparently resides in the oxidized magnetite. AF treatment in higher fields up to 100 mT never isolates stable directions. It was found that samples taken from the same stratigraphic level, but demagnetized either by AF or by thermal methods, had opposite NRM polarity after cleaning. The AF treated samples usually had very scattered demagnetization curves. Often it was difficult to define a characteristic direction (Fig. 6: samples C66-B, C72-B) because the magnetite fraction is obviously very soft. Part of the haematite fraction must also have acquired a secondary magnetization since clearly reversed directions can be observed only during thermal treatment. Well-defined stable NRM directions of either polarity (Fig. 6:

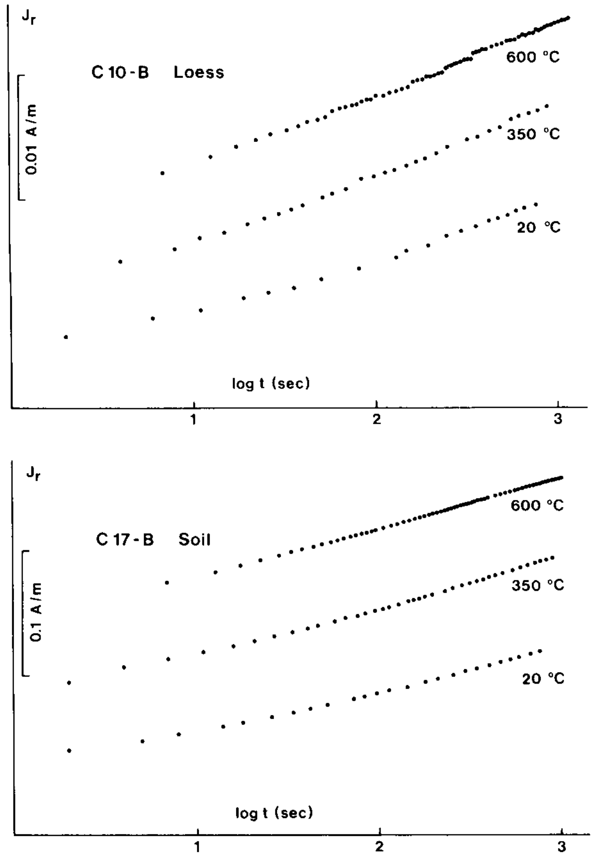

Figure 7. Magnetic viscosity in the NRM state (20°C) and after heating to 350 and 600°C, respectively. The slopes do not change between the heating steps, but in the soil sample they are about 10 times steeper than in the loess sample (note different intensity scales).

C68-A, C60-A) have been obtained after heating to temperature greater than 250–300°C in zero field. This technique was used routinely to establish ChRM vectors.

Fig. 7 illustrates that the secondary NRM component is largely due to viscous remagnetization. The cryogenic magnetometer which is on-line with a HP-1000 minicomputer was cooled to liquid helium temperatures in a small field, thus trapping this field in the measurement chamber. The sample was placed in the magnetometer and the magnetization was recorded in discrete time steps of 5, 10 and 20 s respectively for about 1000 s. The change in remanence was plotted after subtraction of the induced magnetization signal. The resulting slopes of the logarithmic magnetization curves are about 10 times steeper in soils than in loess samples, but cannot always be represented by one viscosity factor for each sample. Sometimes the VRM acquisition contains discontinuities of slope which may be due to the presence of several distinct phases with different activation times (Dunlop 1973; Lowrie & Kent 1978; Walton 1980; Tivey & Johnson 1981). Alternatively the kinks may result from the interaction of at least two viscous remanent magnetization components because they are especially pronounced before any cleaning treatment. In contrast to Tucholka's (1976) observations on VRM in Polish loess, where the viscosity coefficients

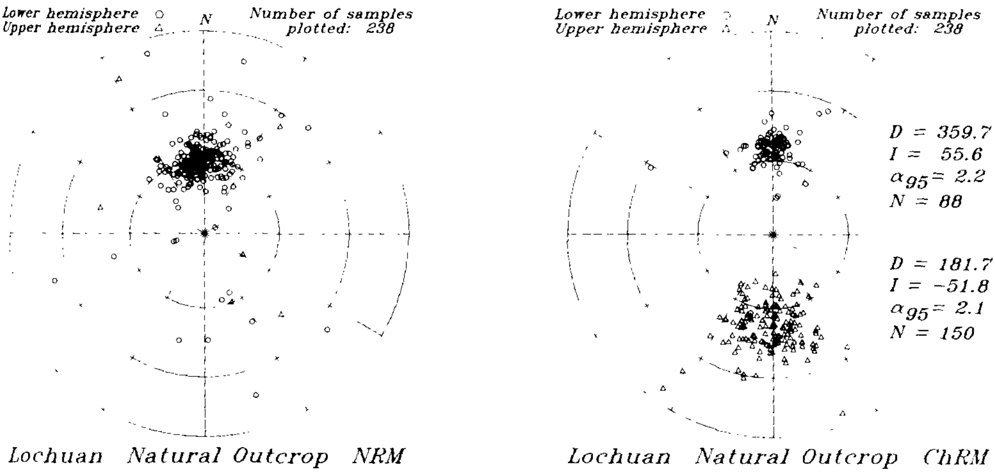

Figure 8. Stereographic projection of (a) NRM directions before and (b) ChRM directions after thermal cleaning. Only the fully oriented samples from the natural outcrop have been plotted.

largely increased after demagnetization at temperatures of 100–300°C, we do not observe significant changes in magnetic viscosity even after thermal demagnetization up to 600°C. Therefore heating obviously does not alter the magnetic properties of Chinese loess sediments. The slopes of the viscous remanence acquisition curves are of the appropriate magnitude to account by extrapolation for the high initial intensities and normal polarity of NRM even in samples with reversed ChRM. Both are caused by viscous remagnetization during the Brunhes epoch.

The stereograms of Fig. 8 summarize the directional NRM behaviour of fully oriented samples taken from the natural outcrop. The NRM directions have been plotted before and after thermal cleaning. Prior to demagnetization nearly all samples are normally polarized due to viscous remagnetization. Demagnetization curves of samples with reversed characteristic magnetization show an increase in remanent intensity between 200 and 300°C due to removal of the normal viscous overprint. These characteristics have been utilized to re-orientate some of the borehole samples which had been orientated erroneously. After cleaning two clearly separated antipodal distributions are observed. Their mean directions agree closely with the axial dipole field direction ($I = 54.9°$) at the sampling locality. The reversed directions are more scattered than the normal directions, probably due to some minor remnants of the secondary overprint.

Magnetostratigraphy

In the earlier studies using AF cleaning (Li *et al.* 1974; An *et al.* 1977; Wang *et al.* 1980), the Malan and the upper part of the Lishi loess were found normally magnetized, and the Wucheng loess predominantly reversely magnetized. The lower part of the Lishi below soil layer S_8 gave inconsistent, poor-quality results. Given the limited available information and the results of thermoluminescence dating indicative of young age, the Lower Lishi was interpreted as originally normally magnetized and belonging to the Brunhes epoch. Consequently the Wucheng with a single intervening normal magnetozone was interpreted as corresponding to the upper Matuyama epoch and the interbedded normal subzone as the Jaramillo polarity subchron.

After thermal demagnetization at 350°C we find the characteristic remanent magnetization of the Lower Lishi below S_8 in both sections investigated to be reversed (Figs 2 and 4). At the base of the Lower Lishi a normal polarity chron has been established which we correlate with the Jaramillo event. Thus the borehole section is subdivided into several clearly defined zones of positive or negative inclination which are separated by sharp transitions (Fig. 2). From top to bottom we associate the polarity changes with:

(1) the Brunhes–Matuyama (B/M) boundary at core depth 53.05 m,
(2) the Jaramillo subchron between 67.30 and 72.50 m,
(3) the Olduvai subchron between 107.40 and 113.10 m,
(4) the Matuyama–Gauss boundary at core depth 137.75 m.

No definite evidence for short-term geomagnetic excursions has yet been found in these data. This may partly be due to several gaps in the sampling column, especially in the upper part of the core. Three samples with anomalous positive inclination (out of a total of 253) do not match the main polarity zonation. They may either be completely remagnetized in the present geomagnetic field or suffer from unrecognized misorientation. Alternatively, they may indeed record short field excursions or short polarity chrons. The two samples near 130 m depth would closely fit the Reunion events. It is questionable whether the normal sample at 60 m depth and the neighbouring shallow inclination samples correspond to the precursor event of the B/M transition as described by Clement, Kent & Opdyke (1982). Enlarged directional scatter of the reversely magnetized samples is found again and is probably due to the incomplete removal of secondary magnetization components. Initial NRM and low field susceptibilities correlate very closely to each other and depend on lithology, the soil horizons being more strongly magnetized than the loess layers.

The natural outcrop has been sampled at four main levels (Fig. 4). Between 49 and 75 m depth three major polarity boundaries could be established which confirm the borehole results:

(1) the Brunhes–Matuyama transition at profile depth 52.30 m,
(2) the Jaramillo subchron between 66.25 and 70.50 m.

The reversal boundaries are slightly offset compared to those of the core due to minor sedimentation differences. The B/M boundary again includes a precursor event, but since two other normally magnetized 'erratic' samples are observed a few metres below, we are again not certain about the significance of this 'event'. The sections between 0 and 12.5 m depth and between 26 and 34 m are both completely normally magnetized. The lowermost section between 115 and 138 m contains a reversal boundary at a depth of 116.25 m which we attribute to the lower boundary of the Olduvai event. As in the borehole the reversal at 137.30 m is interpreted as the Matuyama–Gauss transition. Between these two boundaries all samples collected are reversely magnetized. No evidence of the Reunion events has been found here.

Discussion

THE AGE OF THE LOESS DEPOSITS AT LOCHUAN

The new polarity zonation of Chinese loess deposits (Heller & Liu 1982) the main characteristics of which are now confirmed in the natural outcrop, requires revision of the earlier palaeomagnetic dating results given by Li *et al.* (1974), An *et al.* (1977) and Wang *et al.* (1980). In both sections at Lochuan the B/M boundary is found within the soil layer S_8: (Figs 2 and 4) above the upper silt layer belonging to the Lower Lishi member. The

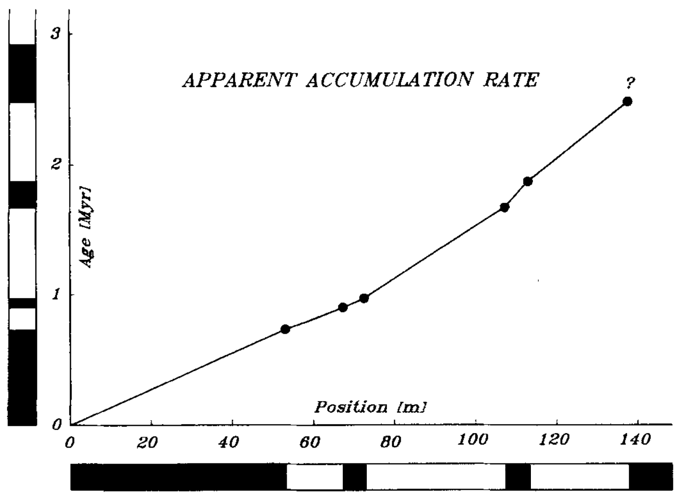

Figure 9. Average apparent accumulation rate in the borehole section at Lochuan.

Wucheng–Lishi transition does not coincide with the beginning of the Brunhes epoch, but is situated below the Jaramillo subchron at a borehole depth of 77.6 m. By linear interpolation between the well-dated (Mankinen & Dalrymple 1979) polarity boundaries an age of about 1.1 Myr is assigned to this lithostratigraphic transition.

The two otherwise rather identical sedimentary sequences of the natural outcrop and the borehole show some lithological differences between soil layer S_8 and the lower silt layer at the top of the Wucheng member. The upper silt layer is reduced in thickness and the soil layers S_9 and S_{10} are stronger developed in the natural outcrop (Fig. 4). The upper boundary of the Jaramillo subchron therefore falls into different lithologies in both profiles, whereas the lower Jaramillo boundary in both sections is placed in the upper part of the loess layer below palaeosoil S_{11}.

Deeper in the borehole the Olduvai subchron has been identified for the first time, implying much greater ages for the entire loess formation than had previously been thought. Converting depth into time and assuming no major sedimentation break between the Upper Pliocene red clay and loess formation (Fig. 9), leads to a maximum age for the start of loess deposition of about 2.45 Myr. The average accumulation rate changes from 4.5 cm kyr^{-1} below the Jaramillo to 7.5 cm kyr^{-1} in the upper part of the sections. Since the loess average density slightly increases below the Jaramillo, we tend to attribute the increasing apparent accumulation rate in the younger sediments to a decrease in compaction due to reduced overburden rather than to increasing sedimentation in the upper part of the profile. A maximum sedimentation age of about 2.4 Myr is also obtained by extrapolation with the lower apparent sedimentation rate beyond the Olduvai subchron without considering the polarity change in the red clay. This suggests that the sedimentation break between the Upper Pliocene red clay and the loess formation may indeed have been of very short duration. Thus the reversal observed in the top part of the Poteh clay most probably reflects the Matuyama–Gauss boundary.

With the earlier maximum loess age of 1.2 Myr given by the Chinese scientists the entire loess formation was placed within the Pleistocene. There is a debate about where to place the Neogene–Quaternary boundary. Chinese researchers prefer the possibility that this transition – if it can be clearly established at all – be situated at the base of the Matuyama epoch. However, according to the resolution of the Joint Meeting of the INQUA Subcommission 1-a and the Working Group of the IGCP Project Nr. 41 (Paris 1980) the Pliocene–Pleistocene boundary stratotype should be defined in the Vrica section (Calabria, Italy) taking into account the first appearance of the early 'cold guest' *Cytheropteron testudo*. This datum is substantially contemporaneous with the first appearance of another 'cold guest', *Arctica islandica* (Pasini & Colalongo 1982).

Magnetostratigraphical work to tie the Vrica section into an age framework related to the geomagnetic polarity history unfortunately allows for alternative interpretations of the Vrica magnetozones. Nakagawa (1981), Obradovich *et al.* (1982) and Tauxe *et al.* (1983) prefer an interpretation which settles the Pliocene–Pleistocene boundary at about 1.6 Myr on top of a normal magnetozone which they assign to represent the Olduvai subchron. This correlation is supported by quantitative studies of selected nannofossil species for which an accurate chronology has been established in open ocean piston cores (Backman, Shackleton & Tauxe 1983). From this evidence the beginning of Chinese loess deposition predates the Neogene–Quaternary boundary so that the loess formation covers the entire Pleistocene as well as some part of the uppermost Pliocene.

GEOMAGNETIC EXCURSIONS DURING THE BRUNHES EPOCH

The durations of polarity intervals through Cenozoic time conform to the first order with a Poissonian distribution (Cox 1968; Lowrie & Kent 1983). This result suggests that a stochastic process controls reversals and that a large number of relatively short duration reversals is expected to occur. Since the first discovery of reversely magnetized Brunhes age basalts (Bonhommet & Babkine 1967) numerous geomagnetic excursions have been reported for the last 730 000 yr. However, the often controversial palaeomagnetic evidence and the imprecise dating has introduced a confusing and frustrating debate amongst palaeomagnetists about the reality and possibility of correlation of these events (Verosub & Banerjee 1977; Verosub 1982).

The closely sampled top section of the natural outcrop at Lochuan (Fig. 4) can provide another check on the possibility of recent excursions, as it covers the last 170 000 yr approximately. This age estimate, derived by linear interpolation between the B/M boundary and the present, is supported by a recent thermoluminescence datum of about 100 000 yr at the profile depth 8–10 m. Excursions postulated as occurring in the last 170 000 yr include the Laschamp (Bonhommet & Babkine 1967), Gothenburg (Moerner, Lanser & Hospers 1971), Erieau (Creer, Anderson & Lewis 1976), Mono Lake (Denham & Cox 1971) and Blake (Smith & Foster 1969) events. Their durations are estimated between 50 and 10 000 yr (Verosub 1982; Heller & Petersen 1982). We have concentrated our detailed sampling especially on the best documented Blake event which is expected in the loess profile at 10 m depth. There is no indication of any directional excursion in the loess profile (Fig. 4). The existence of the Blake or any of the other excursions mentioned cannot be supported. Similarly the section between 26 and 34 m corresponding to an age of 350 000–450 000 yr has strictly normal magnetization directions. Thus the 'β' and 'γ' events in the Gioia Tauro section (Creer, Readman & Jacobs 1980), the Snake River Plain (Champion, Dalrymple & Kuntz 1981) or the Emperor (Ryan 1972) events cannot be confirmed either.

The unconformities and erosional surfaces which separate the lithological loess units, and the different sedimentological development between the upper and lower silt layer in the two profiles, indicate that loess deposition is not a continuous process. Before negating the existence of all the above short excursions we have to consider the possibility of unrecognizable gaps in the loess sequence. However, it seems rather unlikely that there were a large number of sedimentation breaks lasting for several thousand years. Another drawback comes from the timing of the magnetization history. As the characteristic proportion of NRM is carried by haematite of variable grainsize (blocking temperatures between 200 and 650°C), the remanence signal is most probably of chemical origin as suggested by microscopic observation. If the process of haematite growth – in marine limestones the formation of diagenetic haematite is estimated to last for about 100 000 yr (Channell *et al.* 1982) – takes a long time compared to the duration of an excursion, then a majority of haematite particles will be magnetized by the earth field polarity state preceding and succeeding the excursion and the evidence for the excursion itself will be obscured. Such a mechanism might explain the abrupt polarity changes usually observed in the loess at reversal boundaries. The clear evidence for the Jaramillo subchron with a duration of about 100 000 yr, however, suggests that the haematite formation process took at least one order of magnitude less in time than that in marine limestones. Thus polarity events with a duration of about 10 000 yr should be detectable in the loess deposits.

MAGNETIZATION INTENSITIES

There is a direct correlation between NRM intensity, low field susceptibility and loess lithology (Fig. 2). Palaeontological evidence strongly suggests that the lithological variations are caused by climatic changes. Therefore the magnetic intensity variations in loess also depend on climate and do not reflect magnetic field fluctuations as once was suggested for the NRM intensities measured in young deep sea sediment cores (Wollin, Ericson & Ryan 1971). The correlation between climate and geomagnetic field intensity also was questioned seriously by Kent (1982) when he reported a pronounced dependency of NRM intensity and low field susceptibility on deep sea sediment composition which in turn is caused by varying depositional conditions induced by climatic change.

Oxygen isotopes and fluctuations in carbonate dissolution intensities have been used in the past as other high quality recorders of ice volume and climatic changes, respectively. Two palaeomagnetically dated deep sea sediment cores from the central Pacific (Shackleton & Opdyke 1976) and the equatorial Atlantic (van Donk 1976) offer the possibility of a time-based correlation with the magnetic intensity variations in the loess sediments (Fig. 10). For comparison with the V28-239 oxygen isotope record (Shackleton & Opdyke 1976) we have smoothed both data sets with a three-point moving average filter. Both data sets show about the same number of oscillations for the last 2 Myr. Correlation is not straightforward probably because there are some smaller sampling gaps in the loess profile and interpolation between reversal boundaries is only a first-order approximation for the sedimentation history. In both records the variations become especially pronounced after the Jaramillo event. Rather strong glaciations indicated by low susceptibilities (or NRM intensities) occurred about 1.15, 0.85, 0.55, 0.44 and 0.18 Myr ago. Except for the first glaciation which occurred during deposition of the lower silt layer (top of Wucheng member), these events are also clearly visible in the oxygen isotope data and similar absolute ages have been assigned to four glacial maxima during that time period by van Donk (1976).

Shackleton & Opdyke (1976) and van Donk (1976) find the B/M boundary near to their interglacial stage 19. This coincides well with our observations which place the B/M boundary in the top part of palaeosoil S_8. If the soil layer S_2 reflects two closely following warm periods, because it contains a calcareous nodule bed in the middle, then exactly the same number of glaciations are recognized in both the loess and the deep sea record during the Brunhes epoch. Although the amplitudes of the rhythmical susceptibility variations are

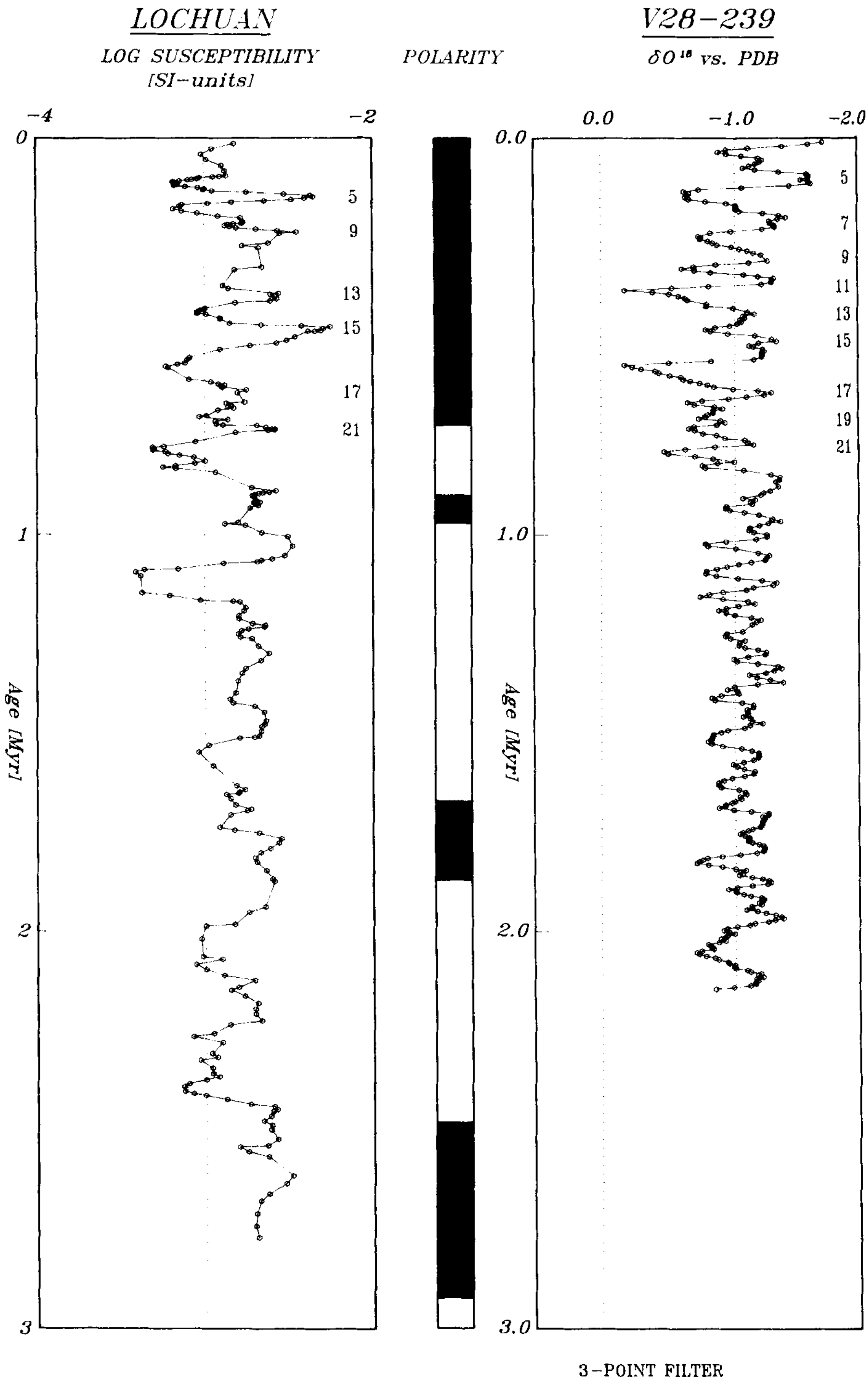

Figure 10. Comparison of the variations of loess susceptibility and oxygen isotope content (redrawn from Shackleton & Opdyke 1976) as a function of time. Numbers along both profiles refer to interglacial stage numbers as introduced by the above authors.

reduced downcore, we can state that 16–18 susceptibility minima occurred since the Olduvai. This result is consistent with the number of glaciations encountered for this time interval in European loess deposits (Kukla 1975; Fink & Kukla 1977). It is also similar to the number of oxygen isotope stages in post-Olduvai deep sea sediment from the Atlantic (van Donk 1976).

Conclusions

The loess sections at Lochuan provide a faithful record of the major polarity changes of the geomagnetic field for the last 2.4 Myr. Loess deposition started at times very close to the Matuyama–Gauss boundary and the Olduvai event has been observed in Chinese loess for the first time. Thus the previous timing of the lithological units constituting the loess formation and – even more important – of the mammalian faunas contained in the loess has now to be reconsidered. The maximum age of Chinese loess formation given in this paper as 2.4 Myr falls near to that reported for European loess and coincides very closely with a pronounced ice volume increase indicated by the oxygen isotope record of deep sea sediments. Recently the first cycle of ice-rafted debris in North Atlantic deep sea sediments due to initiation of major glaciation in northern Europe has been observed to occur at the Matuyama–Gauss boundary (Backman 1983).

Although certain loess sections most probably covering some of the postulated short excursions within the Brunhes epoch such as the Laschamp or Blake events have been sampled in great detail, we were not successful in identifying any events. Either the chemical magnetization process of the loess sediments lasted for a long time, preventing the detection of short-term polarity chrons, or else the existence or world-wide distribution of these short events has to be questioned seriously.

The intensities of NRM, IRM and initial susceptibility are largely controlled by chemical alteration processes and reflect climatic fluctuations during late Pliocene and Pleistocene times. With a more complete and detailed sampling in Chinese loess deposits it will be possible to provide another means of deciphering the climatic history on the northern hemisphere for the last 2.4 Myr.

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