Loess Magnetism

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LOESS MAGNETISM

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Abstract. Loess is a wind-blown Quaternary silt deposit that blankets vast tracts of land and in places reaches thicknesses in excess of 300 m. Over the last decade it has emerged that certain loess sections have recorded the polarity history of the geomagnetic field and now provide essentially continuous magnetostratigraphic archives covering the last 2–3 m.y. Indeed, it is the chronology provided by the magnetic polarity signature itself that was largely responsible for establishing the timing of the initiation of loess accumulation, particularly in the celebrated Chinese loess plateau, where a starting date close to the Gauss-Matuyama chron boundary (2.6 Ma) is now firmly established. This coincides with a widely documented global climatic shift and accelerated uplift of the Tibetan plateau. Many loess sections contain fossil soils (paleosols) that bear witness to warmer and wetter climatic conditions corresponding to interglacial periods in contrast to the cold, arid environments in which pristine loess accumulated and which correspond to glacial intervals. The resulting sequences of alternating loess and paleosols also manifest themselves magnetically, in this case in terms of susceptibility changes, entirely distinct from the remanence characteristics, which encode the geomagnetic polarity. The susceptibility time series obtained from localities in Alaska and China correlate remarkably well with the oceanic oxygen isotope signal and yield spectral power estimates in agreement with those predicted by the astronomical (Milankovitch) theory of ice ages. Comparison of susceptibility patterns with corresponding profiles of ^{10}Be concentration in loess allows major changes in rainfall to be estimated. In China, for example, data spanning the last 130 kyr (corresponding to oxygen isotope stages 1–5) indicate that paleoprecipitation was almost halved (from ~540 to ~310 mm yr^{-1}) as the warm interglacial during which paleosol S_1 formed gave way to the following glacial interval in which loess layer L_1 accumulated. It has also been found that increased amounts of continent-derived dust delivered to the deep ocean correlate with loess formation and thereby permit certain broad features of atmospheric circulation (paleowinds) to be worked out. Debate continues over the actual mechanism by which magnetic susceptibility becomes a climate proxy. The current consensus is that some form of in situ process must be responsible, at least in part. Detailed laboratory investigations, both on whole samples and on magnetic extracts, indicate that the enhancement observed in midlatitude weathered loess and paleosols is largely due to a magnetically "soft" mineral which is either magnetite (Fe_3O_4) or maghemite (γ-Fe_2O_3). Experimental evidence is accumulating that tiny (<100 nm) ferromagnetic particles probably generated by the activity of magnetotactic bacteria in the soil are responsible.

INTRODUCTION

Our title prompts two obvious questions: What is loess, and why study its magnetism? Loess has been defined in many different ways by many different authors, but for present purposes it can be regarded essentially as a terrestrial wind-blown silt deposit. The word loess (German "Löß") was apparently coined in the geological literature of the early nineteenth century as a name for loose silty deposits along the Rhine near Heidelberg and was brought into English by Sir Charles Lyell following his visits to the valleys of the Rhine (1833) and the Mississippi (1845–1846). In China, where loess is called "huangtu" (yellow earth), the relation of loess with wind-blown dust was already known 2 millennia ago [*Liu et al.*, 1985a]. The eolian origin of loess was not at first recognized in the western world but became popular after the seminal work of *von Richthofen* [1882] describing the magnificent loess deposits in China. Although a great deal of the world's loess is undoubtedly of primary eolian origin, a significant part has suffered some degree of reworking and subsequent redeposition. Furthermore, both primary and reworked (sometimes called loesslike) deposits are, in many cases, modified by weathering, soil formation, and diagenesis. All these categories are important, and we discuss results from each of them.

Loess and loesslike deposits cover about 10% of the world's land surface and are concentrated in the temperate climate zones (Figure 1). They are distributed

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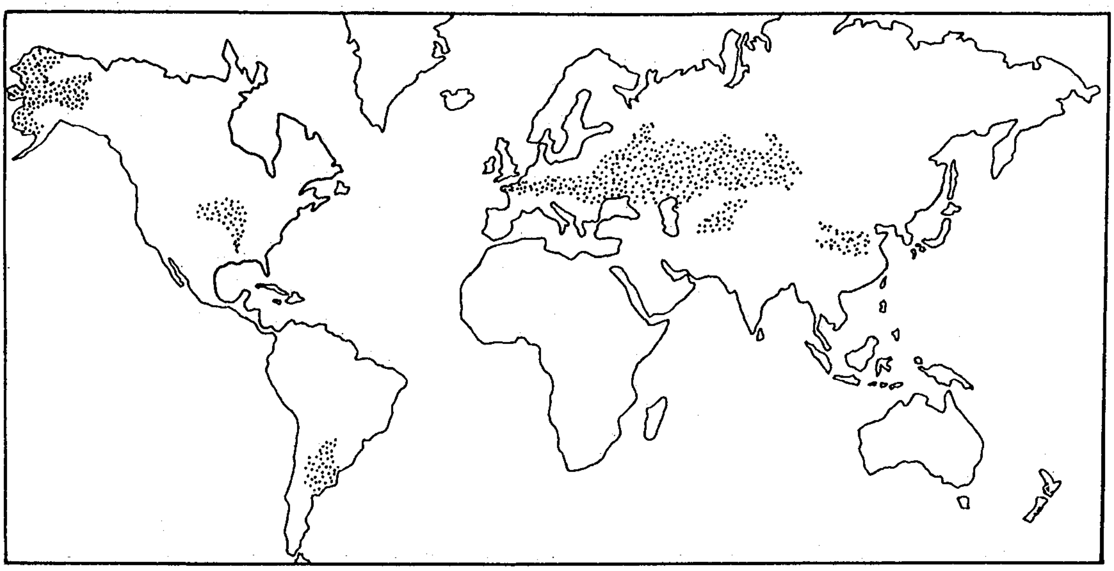

Figure 1. The global distribution of major loess occurrences [after *Catt*, 1986; *Pye*, 1987]. Because loess thins out often in a nonmappable manner, boundaries of loess areas (dotted) are omitted, and numerous smaller and thinner occurrences are not shown.

across former periglacial regions adjacent to glacial till areas and semiarid desert margins. As *Liu et al.* [1985a] point out, the history of mankind is closely linked to the distribution of loess, as is the modern economy, since highly industrial as well as agricultural areas are situated in the loess belt.

Most commonly, loess is a homogeneous and highly porous sediment devoid of stratification or lamination. It is buff to yellow in color and may darken with increasing degree of weathering. The grain size distribution shows a typical mode in the range 20–40 μm [*Pye*, 1987], but more sandy or more clayey loess is not unusual either. The major rock-forming minerals in loess are quartz and feldspars (up to 90%). The feldspars may be converted to and replaced by clay minerals, especially smectite and illite during weathering [*Bronger and Heinkele*, 1990]. The heavy mineral fraction almost always contains ferromagnetic minerals such as magnetite, maghemite, hematite, and goethite [*Heller and Liu*, 1984].

The loess material is thought to be formed originally by two main processes. On the one hand, glacial grinding, frost action, and fluvioglacial abrasion [*Smalley*, 1966] may provide the silt material which can be transported later by strong and turbulent winds over large distances (Figure 2). Significant amounts of silt, on the other hand, may also be produced in cold and dry deserts by frost action and salt weathering [*Derbyshire*, 1983]. Wind transportation certainly has occurred in multiple steps during cold and stormy climate periods, so that the loess found in its present position has undergone frequent episodes of erosion and redeposition.

More humid and warmer interglacial or interstadial times provoked weathering and alteration of loess into soils that being now covered by more recent loess are called paleosols (for details of the processes of soil formation or pedogenesis, see, for instance, *Catt and Weir* [1981]). In many loess regions the changing paleoclimatic conditions lead to an alternating succession of paleosols that represent more humid and warmer conditions and loess beds that formed during colder and more arid climate stages. The evidence for these climate changes is diverse: paleontological evidence sees changing mollusc assemblages [e.g., *Rousseau and Puisségur*, 1990], calcium carbonate is depleted and Fe_2O_3/FeO increased in paleosols [*Wen et al.*, 1982; *Liu et al.*, 1985b], clay minerals are enriched by weathering processes in paleosols [*Bronger and Heinkele*, 1990], and investigations of micromorphological structures of the paleosols enable differentiation of variable interglacial vegetation stages [*Bronger and Heinkele*, 1989].

Over the last decade or so it has emerged that many loess deposits around the world have preserved remarkable records of several million years of both climatic and magnetic history. These natural archives are currently under intensive study and are providing important data for the alternating polarity states of the Earth's magnetic field and for the waxing and waning of the ice ages. The magnetic properties of loess deposits seem to carry high-resolution information of both the geomagnetic and climatic histories and have thus come to be a central concern to a broad spectrum of research geoscientists and of more than passing interest to a very wide readership.

The magnetic properties germane to the present discussion are remanence and susceptibility. Other factors creep in later on, but these two are of paramount importance throughout, and it is essential to

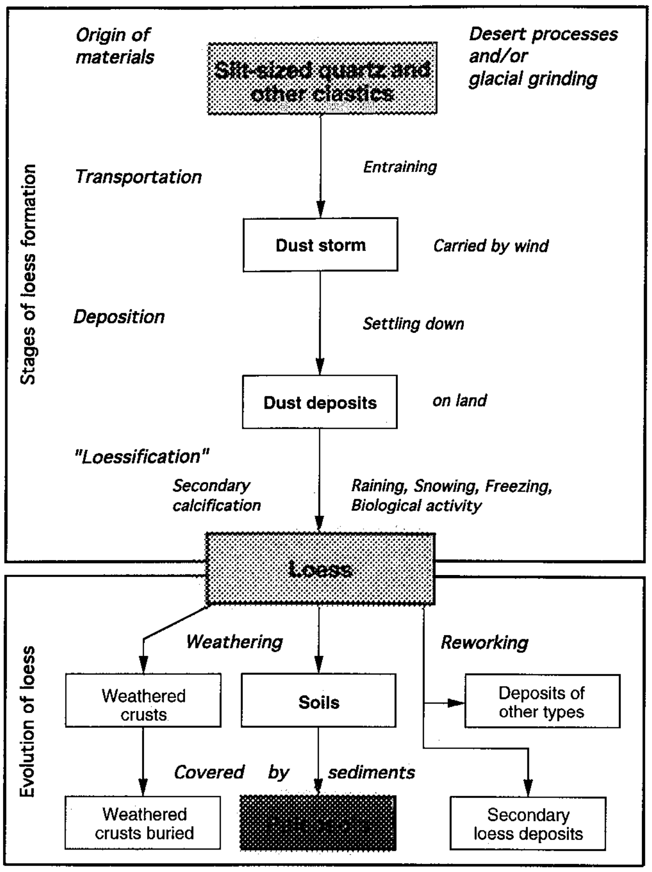

Figure 2. Schematic diagram of the formation of loess deposits which includes the formation of loess-sized particles and their transport, deposition, and postdepositional modification (adapted from *Liu et al.* [1985a]).

keep them distinct. Remanence is the technical name given to the "permanent" magnetization acquired at (or some time after) deposition; it is responsible for recording the geomagnetic polarity changes. Susceptibility is the parameter describing how "magnetizable" a material is when subjected to a known external magnetic field; essentially, it measures the amount of magnetic material present and thus is the means by which paleoclimatic information comes to be stored in loess strata. Remanence is a vector quantity; susceptibility is a scalar. The latter is readily determined on commercially available meters which are handy enough to be taken into the field; measurements are quick and easy. Remanence, on the other hand, is always studied in the laboratory; it requires accurately oriented samples and much more sophisticated spinner or cryogenic magnetometers. For details on magnetic properties, field and laboratory procedures, and broad environmental applications, refer to the excellent texts by *O'Reilly* [1984], *Collinson* [1983], and *Thompson and Oldfield* [1986], respectively.

As indicated above, there is a wide variety of loess deposits throughout the world; the major occurrences are indicated in Figure 1. The thickest sections are found in China and central Asia, where thicknesses up to 300 m are known. In the Danube basin, in Argentina, and in parts of North America, thicknesses in excess of 60 m are reported, but elsewhere, 30 m is more typical. Even when attention is restricted to work concerning magnetic properties, the published literature is enormous and involves several languages. It is not our intention therefore to provide an exhaustive, detailed summary of every single locality from which results are known. Rather, our aim is to bring together a representative sample of the most important results from each of the general areas indicated in Figure 1. We admit at the outset a deliberate bias toward China, partly because these are the results with which we are most familiar but also because it is in the celebrated Chinese loess plateau that some of the most striking outcrops are found and from it that some of the most significant scientific findings have been forthcoming.

GLOSSARY

Anhysteretic remanent magnetization (ARM): remanent magnetization acquired in a small, steady field superimposed by a regularly decreasing alternating field (AF).

Eemian interglacial: last interglacial, between 70,000 and 128,000 years B.P.

Gley: soil of greyish color due to the reduction of ferric oxides and hydrated oxides in the presence of organic matter caused by waterlogging.

Interglacial: interval between two glacial periods characterized by warmer climate, when ice was generally melting.

Interstadial: relatively short warm interval between two glacial intervals within a major glacial period.

Isothermal remanent magnetization (IRM): remanent magnetization acquired at room temperature in progressively higher steady fields.

Solifluction: slow flowing from higher to lower ground of masses of waste saturated with water, caused by snowmelt or rain.

Superparamagnetism: phenomenon occurring in small ferromagnetic grains with short relaxation times. Remanent magnetization is unstable and decays very rapidly to zero after removal of the magnetizing field.

Verwey transition: temperature at which a crystallographic transition from cubic to orthorhombic occurs in magnetite. This is connected with a loss of remanent magnetization in magnetite grains, the magnetic energy of which is controlled by magnetocrystalline energy (in general multidomain grains).

Virtual geomagnetic pole (VGP): pole position resulting from a single paleomagnetic direction, calculated by assuming a dipolar geomagnetic field.

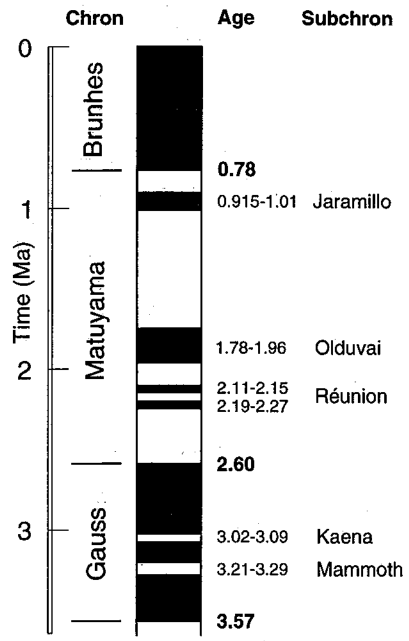

Figure 3. Polarity timescale of the youngest three polarity chrons (Gauss to Brunhes). Age of reversal boundaries according to *Spell and McDougall* [1992] and *McDougall et al.* [1992].

MAGNETOSTRATIGRAPHIC POLARITY ZONATION

Magnetostratigraphic polarity dating of sediments simply correlates observed polarity patterns with the well-established geomagnetic polarity timescale [*Cande and Kent*, 1992]. Since loess has been deposited throughout the Quaternary, such a comparison is expected to be made for the last few million years only. The lower boundary of the Quaternary itself is still debated; it is placed either at the upper boundary of the Olduvai subchron as proposed by *Nakagawa* [1981] or at the Gauss-Matuyama transition as preferred by Chinese scientists [*Liu et al.*, 1985a]. Regardless of these problems, the polarity pattern for the last 3.5 m.y. will be sufficient for age determination within the loess (Figure 3). During this interval, three chrons (the normal Gauss, the reversed Matuyama, and the normal Brunhes) determine the main polarity pattern. Within these chronozones, several subchrons such as the Olduvai or Jaramillo intervene. These fairly young chrons, and also the main chron boundaries, have recently been redated radiometrically [*Spell and McDougall*, 1992; *McDougall et al.*, 1992] and astronomically on the basis of Milankovitch cyclicities [*Johnson*, 1982; *Shackleton et al.*, 1990; *Bassinot et al.*, 1994]. Figure 3 does not show any of the very short-lived polarity subchrons such as the Blake, which occurred at approximately 115 ka [*Champion et al.*, 1988; *Zhu et al.*, 1994a]. The temporal and spatial occurrence of these is usually not well documented, and the reality of many of them remains debatable. We will, however, discuss possible occurrences in the various loess profiles.

China

Despite several early attempts [*Li et al.*, 1974; *An et al.*, 1977], it was little more than a decade ago that a convincing magnetic polarity zonation for the Chinese loess plateau was first established [*Heller and Liu*, 1982]. This seminal result, based on 231 samples taken from a borehole in the vicinity of Luochuan (35.8°N, 109.2°E), is illustrated in Figure 4. The boundary between the Brunhes and Matuyama chrons (M-B) occurs at a depth of 53.05 m, with the Jaramillo and Olduvai subchrons spanning the intervals 67.30–72.50

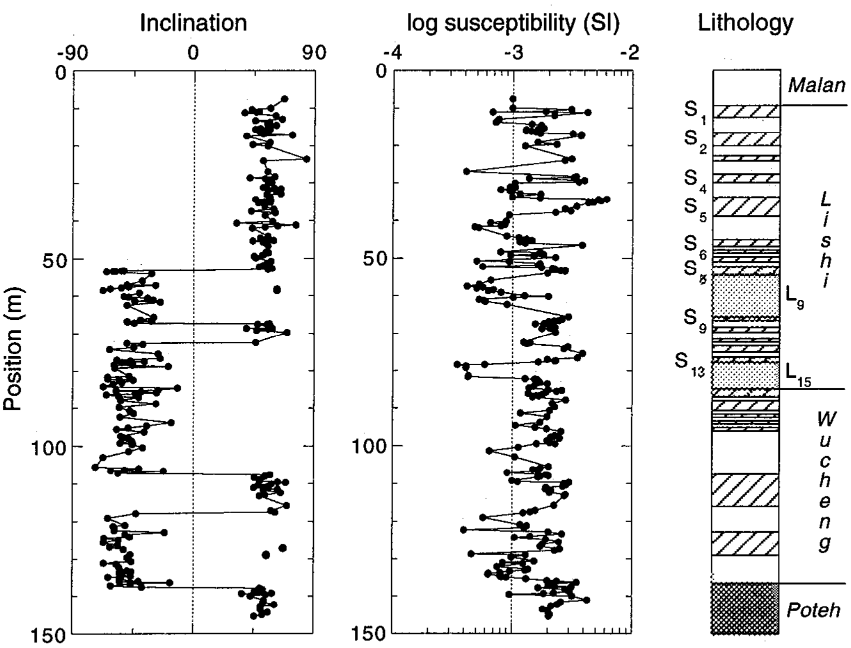

Figure 4. Magnetostratigraphy of the borehole at Luochuan. The lithological units are named Malan, Lishi, and Wucheng loess [cf. *Liu and Chang*, 1964]. The Malan loess contains mammalian fossils typical of late Pleistocene age. Elements of Zhoukoudian faunas (equivalent to the European Oldenburgian) have been found in the Lishi member. The Wucheng loess is characterized by the Villafranchian Nihewan fauna considered to be early Pleistocene. The loess units are preceded by the Poteh red clay which corresponds to the uppermost Gauss chron and contains upper Pliocene mammalian remains [*Teilhard de Chardin and Young*, 1930]. In the lithology column, loess layers (L) are white or stippled if silty; paleosols (S) are hatched. Redrawn from *Heller and Liu* [1982, 1984].

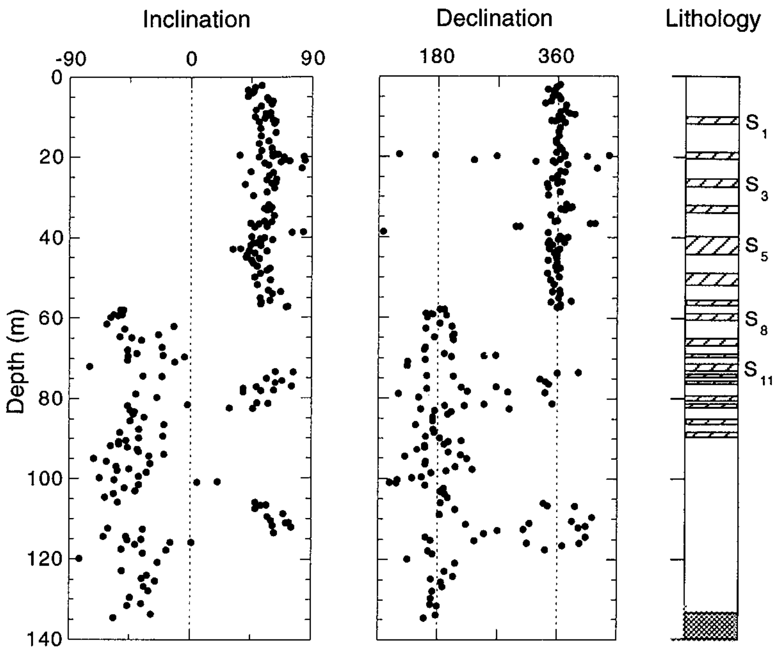

Figure 5. Magnetostratigraphy at Luochuan after thermal cleaning of natural remanent magnetization (NRM) at 300°C after *Torii et al.* [1984].

m and 107.40–117.70 m, respectively. The base of the loess itself falls at 136 m, below which a widespread formation known as the Red Clay is found. Prior to the publication of these borehole data, loess deposition in China was generally thought to have commenced no earlier than 1.2 Ma, but now for the first time the Olduvai subchron was recognized, and linear extrapolation implied a minimum starting age of deposition of ~2.4 Ma. At more or less the same time that the borehole data became available, a joint Japanese-Chinese team were investigating the magnetic remanence of 200 block samples collected from natural outcrops within a few kilometers of Luochuan. From a total thickness of 134 m they obtained a magnetostratigraphic zonation very similar to that deduced from the borehole data [*Torii et al.*, 1984], as can be seen in Figure 5.

Heller and Liu [1982] also investigated samples from below the base of the loess in the underlying Red Clay and found a return to normal polarity (Figure 4). The suggestions were therefore made that sedimentation was continuous between the loess and the Red Clay and that the normal polarity thus represents the top of the Gauss chron. Subsequently, *Heller and Liu* [1984] confirmed this suggestion using new samples collected from a natural outcrop about 3 km from the borehole site.

Following these early reports, magnetic studies spread rapidly westward and southward from Luochuan, so that complete profiles are now available from natural outcrops at Lanzhou [*Burbank and Li*, 1985; *Rolph et al.*, 1989], Xifeng [*Liu et al.*, 1987, 1988], Baoji [*Rutter et al.*, 1990, 1991], and Lantian [*Zheng et al.*, 1992]. Broadly speaking, these new data confirm the original picture, although certain points require attention. With reference to Figure 6, it can be seen that the Baoji section (34.4°N, 107.1°E) provides a close parallel to the Luochuan interpretation. On geological grounds, *Rutter et al.* [1991] argue that it should be regarded as the type pedostratigraphic or soil stratigraphic section for the loess plateau. On the whole, the Baoji magnetic profile (162 m, 483 samples) provides a very clear magnetic zonation, but a few data points apparently yield the "wrong" polarity, particularly just above the Jaramillo and just below the Olduvai. These may represent real geomagnetic features, but it must be noted that they all possess normal polarity and may simply be overprinted samples that have not responded to laboratory cleaning.

At Xifeng (35.7°N, 107.6°E), *Liu et al.* [1987, 1988] obtained a polarity sequence exhibiting clear boundaries at the top and bottom of the Matuyama chron at depths of 71 and 180 m. They also found convincing evidence for the Olduvai subchron, but near the expected position of the Jaramillo subchron the profile is confused, with 10 or more apparent inversions over a 30-m interval. Thus, although the general picture is not in doubt, there are obviously some imperfections, either with the recording fidelity of (some) loess or with the procedures commonly in use.

Zheng et al. [1992] report a magnetostratigraphic profile from a 133.5-m-thick section of loess at Lantian, 20 km east of Xian (34.2°N, 109.2°E), from which 320 blocks were collected. The resulting interpretation puts the top and bottom of the Matuyama chron at depths of about 50 m and 132 m, respectively, with clearly revealed Jaramillo and Olduvai subchrons in between. Short intervals of normal polarity are again found above the Jaramillo and below the Olduvai, and an additional feature appears some 10 m below the Jaramillo.

In the far west of the loess plateau, conflicting

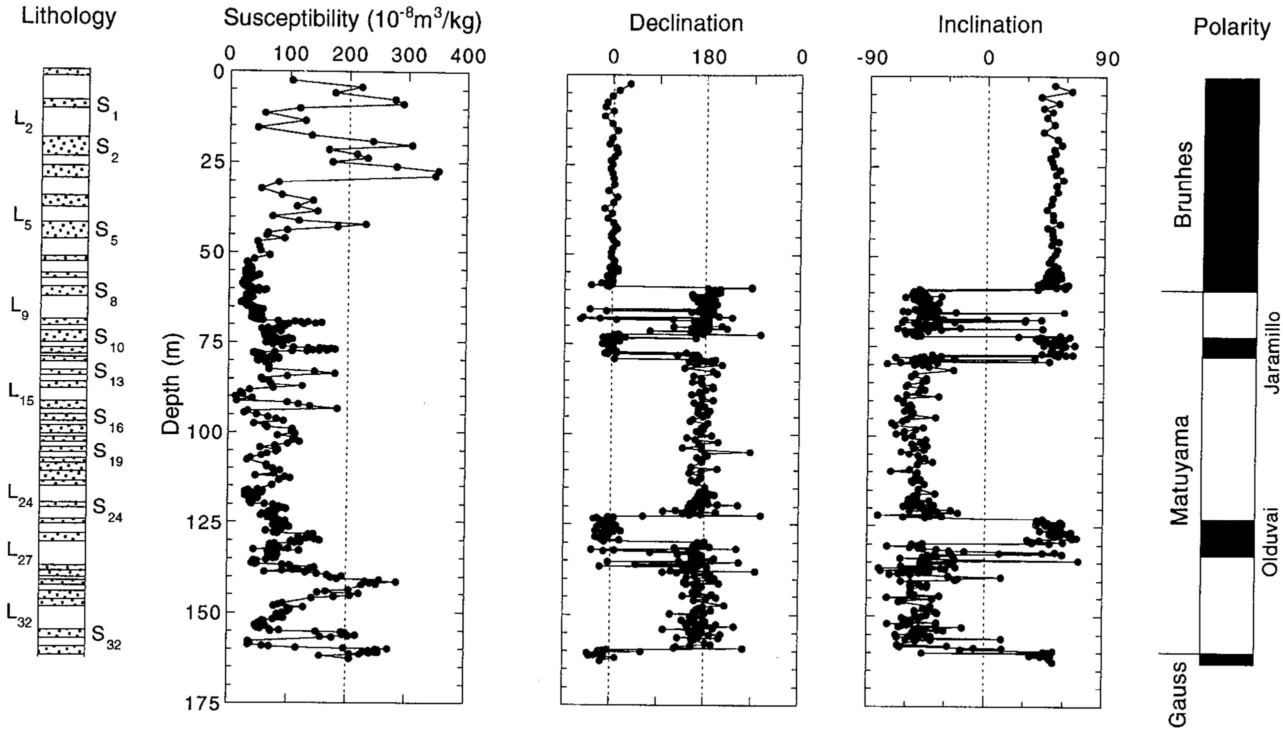

Figure 6. Loess/paleosol lithology and susceptibility at Baoji and polarity zonation defined by declination and inclination of 482 alternating field (AF) and thermally cleaned oriented samples [from *Rutter et al.*, 1991].

results have been obtained from outcrops near Lanzhou. *Burbank and Li* [1985] deduced that loess accumulation commenced much later in this area than at Luochuan despite the greatly enhanced thickness of the sequence. On the basis of 40 sampling horizons spanning some 330 m they suggest the interpretation shown on the left side of Figure 7, in which the uppermost ~180 m represents the Brunhes chron. Two normal polarity horizons lower down are assigned to the Jaramillo subchron, and the base of the section is

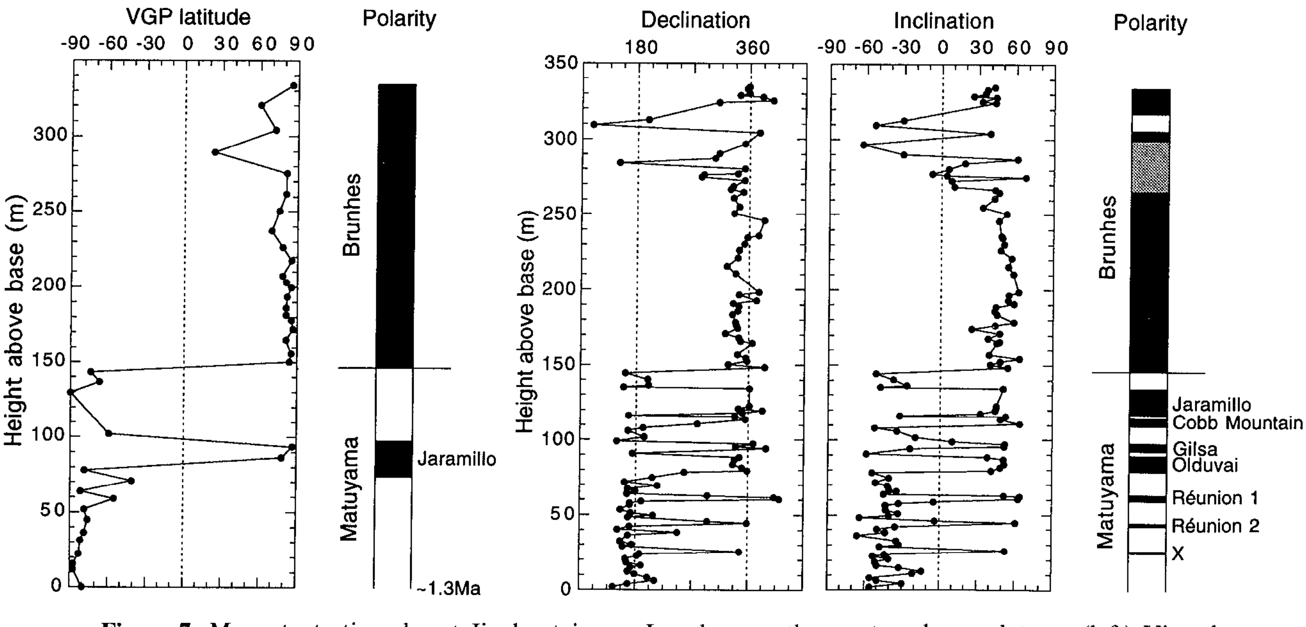

Figure 7. Magnetostratigraphy at Jiuzhoutai near Lanzhou on the western loess plateau. (left) Virtual geomagnetic pole (VGP) results and interpretation by *Burbank and Li* [1985]. (right) Declination and inclination data and interpretation by *Rolph et al.* [1989]. Short-lived subchrons in the Matuyama chron according to *Mankinen and Dalrymple* [1979].

thus extrapolated to have an age of about 1.3 Ma. This markedly different chronology from that obtained at sites more centrally situated in the loess plateau obviously has important repercussions for Pleistocene stratigraphy in China as a whole. Consideration must be given to the results reported by *Rolph et al.* [1989] from the same area and consisting of a composite section comprising 116 sampled horizons. Their results are summarized on the right side of Figure 7 for comparison with the Burbank and Li data. The newer results imply a more complicated magnetic zonation, and Rolph et al. admit to some difficulty in correlating to the standard polarity timescale. Their preferred interpretation requires that the normal polarities occurring about 60 m below the Matuyama-Brunhes boundary represent the Olduvai subchron, not the Jaramillo as Burbank and Li propose. They then correlate the six "extra" normal intervals with the standard timescale as shown in Figure 7. This leads to an age of ~2.4 Ma for the basal loess in this area, in much closer agreement with the more central sites. However, this interpretation requires a rather contrived sedimentation history in order to accommodate the observed thicknesses; for times prior to ~1.9 Ma the apparent accumulation rate is 12.5 cm kyr^{-1}; from then until ~0.9 Ma the rate is 4.5 cm kyr^{-1}, and thereafter it increases dramatically to 26 cm kyr^{-1}. It remains to be seen if further magnetic work and/or geological information can reconcile these two divergent views. Meanwhile, it is prudent to note that the sections near Lanzhou form very steep slopes which are prone to landslides (E. Derbyshire, personal communication, 1994), which may well confuse the magnetostratigraphic picture.

Several attempts have failed to identify records of short geomagnetic polarity changes in the loess [*Heller and Liu*, 1984; *Liu et al.*, 1987; *Zheng et al.*, 1992], but evidence for the reality of such features recorded in the Chinese loess has been recently reported. *Zhu et al.* [1994a] investigated the nearly 6-m-thick paleosol S_1 near Xining, Qinghai (101.8°E, 36.6°N), in one of the huge 320-m-thick outcrops on the western plateau. Over 700 samples were collected from three profiles across this pedocomplex, which has been dated by thermoluminescence and by stratigraphic comparison with the marine oxygen isotope timescale between 72 and 130 ka. They find evidence of a tripartite reversal pattern over a stratigraphic thickness of about 60 cm, which they attribute to the Blake episode [*Smith and Foster*, 1969]. In the loess section investigated, estimates of the deposition rate based on the magnetozone boundaries imply that the Blake lasted from 117.1 ± 1.2 to 111.8 ± 1.0 ka.

Central Asia

A great deal of paleomagnetic work has been carried out on the central Asian loess deposits, which show broadly similar properties to those of the Chinese loess plateau. They are as widespread as in China (Figure 1) and approach thicknesses well over 200 m with up to 35 intervening paleosols. In the younger part the uppermost six or seven loess units often approach thicknesses well over 10 m like the late Pleistocene Malan loess in China. Individual loess beds below the M-B boundary are much reduced in thickness (1–4 m) and are similar to that of the paleosols. According to magnetostratigraphic evidence the maximum deposition age may exceed 2 Ma. The most detailed studies have been made in Uzbekistan in the area around Tashkent and in Tajikistan, where the Mesozoic-Cenozoic infill of the Tajik depression is capped by loess sediments of lower, middle, or upper Pleistocene age [*Pye*, 1987].

Some of the most complete sections are situated 100–200 km east to southeast of Dushanbe, the Tajik capital city, at Karamaydan and Chashmanigar [*Dodonov et al.*, 1977; *Dodonov*, 1991]. The magnetic polarity stratigraphy of these sections and many others that are not so continuous or are less well exposed has been investigated by *Gamov and Pen'kov* [1977]. Their results, however, have been presented only in the form of interpreted magnetic polarity columns without documenting the actual data or reporting details of experimental analysis performed in Pen'kov's paleomagnetic laboratory at Dushanbe. *Pen'kov and Gamov* [1977] applied prolonged thermal demagnetization for several hours between 140°C and 180°C to 30% of a very large number of samples (>15,000). The remainder were stored in field-free space in order to reduce or eliminate secondary magnetic overprints.

The M-B boundary in the 250-m-thick loess-paleosol sequence at Karamaydan (37.6°N, 69.1°E) is found at a depth of ~117 m directly above the tenth pedocomplex. The beginning of the Jaramillo subchron occurs at a depth of 156 m, and the Olduvai subchron is found between 192 and 201 m. The Gauss-Matuyama (G-M) transition lies at 238 m depth. The magnetostratigraphy is characterized by many short excursions, which are observed mainly in the Matuyama period but also in the Brunhes. They are interpreted as real geomagnetic features, and stratigraphic correlation between sections has been attempted especially for the Brunhes excursions [*Lazarenko et al.*, 1991]. Slightly diverging results have been reported recently for the Karamaydan section by *Forster and Heller* [1994]. They report a composite profile which yields an overall thickness of continuous loess deposition of ~100 m only. Further continuation to depth was not possible because sedimentation continuity could not be guaranteed and natural outcrop in a parallel valley may not exceed another 20 m. Paleomagnetic analysis of nearly 250 samples confirmed the position of the M-B boundary between the ninth and tenth pedocomplexes but at a smaller depth of ~85 m. No evidence was found for the presence of the

Jaramillo subchron, so that the section covers a total time span of <1 m.y.

On the basis of Pen'kov's work, *Dodonov* [1986] describes similar relations for Chashmanigar, where, however, the G-M boundary is not detected. The 174-m-thick profile has the M-B boundary at a depth of 80 m, again above the tenth pedocomplex; the Jaramillo subchron is situated between 97 and 107 m, and the Olduvai is between 160 and 167 m at the level of the 35th paleosol. Both sections, Karamaydan and Chashmanigar, have in common that the M-B boundary is observed in the same stratigraphic horizon, below pedocomplex 9.

The loess in Uzbekistan is not as thick as that in south Tajikistan, but it reaches some 90 m at Orkutsai (also called Charvak), which is located about 100 km east of Tashkent [*Lazarenko*, 1984; *Dodonov*, 1991]. The paleomagnetic data provide evidence that the oldest preserved loess dates from the Jaramillo subchron between 83 and 89 m depth, whereas the M-B boundary is positioned between pedocomplexes 9 and 10 as in Tajikistan, at a depth of 60 m.

Europe

A belt of well-documented loess deposition extends in the periglacial realm between the Alpine glaciers and the former Scandinavian ice sheets. Along the Danube river and its tributaries and along the Rhine river, several continuous or semicontinuous sections have been studied paleomagnetically.

The geologically younger brickyard exposure at Red Hill (Czech Červený Kopec) in Brno, Czech Republic, and the older one at Krems, Austria, which are among the earliest sections studied and dated paleomagnetically, have been interpreted magnetostratigraphically to overlap in the age range around the Matuyama-Brunhes boundary [*Kukla*, 1975; *Kocí and Sibrava*, 1976; *Fink and Kukla*, 1977]. The two sections constitute an 80-m-thick series of alternating loess and paleosol beds where the M-B boundary is observed at an approximate depth of 65 m below pedocomplex 9. Apparently, the sedimentation rate changes abruptly below this boundary so that the Jaramillo subchron forms a 1-m-thick normal zone at about 67-m depth and the Olduvai subchron is found in the lowermost 5 m. The M-B boundary is defined by a clear sign change of the inclination value, whereas declination values are largely scattered. *Kocí* [1990] finds the Jaramillo subchron preceded by two short, normal polarity episodes which appear, however, doubtful in view of the rapid and irregular natural remanent magnetization (NRM) directional changes observed in this interval of the Red Hill section. Sediments at Brno that correspond to the last interglacial carry southerly declinations but maintain positive inclinations and have been interpreted as correlative with the Blake event [*Kukla and Kocí*, 1972].

The well-documented "Young" and "Old" loess successions along the Danube river in Hungary began to accumulate just after the Jaramillo subchron. The polarity data presented by *Pécsi* [1985] show the M-B boundary in almost the same stratigraphical level at a depth of ~46 m near the very bottom of the profile at Paks (18.9°E, 46.6°N) and at 42 m in the profile at Dunaföldvár (18.9°E, 46.8°N). The data from Paks are based on a detailed study of nearly 100 paleomagnetic samples by *Márton* [1979], who demonstrated the reliability of the NRM directions by careful AF demagnetization. The Jaramillo subchron, however, was detected in the nonloessic sedimentary substratum only. In the 60-m-deep Posta valley borehole near Pécs (18.2°E, 46°N), which penetrates an older loess/paleosol succession, the polarity of 10 samples has been measured and interpreted as documenting the time span from the Brunhes into the Gilbert polarity epoch.

The clay pit at Kärlich near Bonn, Germany, contains six or seven alternating loess/paleosol beds, which according to measurements of 26 samples are normally magnetized throughout [*Kocí and Sibrava*, 1976; *Brunnacker and Boenigk*, 1976]. The M-B boundary is located in a gravel horizon just underneath the lowermost loess bed. Interlayered tuff horizons have been dated at about 450 ka. Therefore it was argued that loess accumulation at Kärlich commenced shortly after the M-B boundary [*Brunnacker and Boenigk*, 1976; *Wiegank*, 1990]. The older part of the thin loess blanket in northern France (Normandy) is reversely magnetized, whereas the younger series apparently belongs to the Brunhes epoch [*Lautridou*, 1979]. No detailed paleomagnetic data have yet been published, and even the position of the M-B boundary is unknown.

In Poland, loess sedimentation probably started after the M-B boundary, but to our knowledge no coherent, well-documented magnetostratigraphic profiles have been reported [cf. *Wiegank*, 1990]. However, *Tucholka* [1977] has investigated in great detail the youngest part of four approximately 15-m-thick loess sections in Poland. He draws attention to the occurrence of an excursion with anomalously low and sometimes even negative inclination, which occurs in a loess horizon postdating the Eemian paleosol. Because of the stratigraphic position and since the inclination excursion in one profile has a bipartite pattern [*Creer et al.*, 1980], it may be considered as documenting the Blake polarity subchron.

Numerous sections have been studied by Russian scientists in the loess wide spread over the east European platform. Their results have been published almost exclusively in Russian and usually have been inaccessible to us. The paper describing the magnetic results from the Novaya Etuliya section [*Faustoff et al.*, 1986] has been translated by E. Virina(Department of Geography, Moscow State University). This section, which is situated in the southern part of Moldavia (29.9°E, 45.9°N), has a thickness of about 35 m and consists of an alternating loess/paleosol series down to 24 m. The lowermost 11 m consist of variegated silts and clays with intervening paleosols. The Brunhes part of the section covers an apparently complete sequence of nine paleosols and corresponding loess beds similar to the paleosol number in China or Tajikistan. The M-B boundary is found at a depth of 14.5 m, indicating sedimentation rates on the Russian platform that are typically low compared with those of China or central Asia. Within the underlying section representing the reversed Matuyama period, two normal polarity intervals (each 2 m in thickness and located in a paleosol horizon) have been observed that may be interpreted as Jaramillo and Olduvai subchrons. The latter subchron terminates the loess succession at 24 m. The loess formations in the foreland hills of the Precaucasus region attain larger thicknesses. The Otkaznoe section near the city of Zelenokumsk (43.5°E, 44.2°N) consists of two loess units with a total thickness of about 140 m [*Virina et al.*, 1990]. The lower 45-m-thick unit contains a loess/paleosol series with three soils and loess beds that are reversely magnetized and belong to the Matuyama period. The Jaramillo subchron has been recognized between 94 and 104 m. The upper series includes five pedocomplexes and six loess horizons with the M-B boundary at a depth of 72 m.

North America

As well as the loess deposits known in the Mississippi valley since the time of Lyell, there are widespread deposits in interior Alaska, some of which exceed thicknesses of 50 m [*Péwé*, 1955]. Recently, detailed paleomagnetic results have been obtained from a number of exposures in the general vicinity of Fairbanks [*Westgate et al.*, 1990]. The most important data are from the so-called Gold Hill loess at their locality 6 (147.9°W, 64.9°N). A total thickness of 32 m was sampled at 10-cm vertical intervals, and five normal/reversed couplets were found (Figure 8). The polarity zones implied vary in thickness from <1 m to >12 m, and there are three recognized unconformities. However, there are present a number of tephra layers which are of great diagnostic value throughout northwestern North America, and these help establish the reliable magnetostratigraphy indicated. This is a further example where paleomagnetic study has provided important chronological control; in this case a complete reassessment of the antiquity of loess deposition in Alaska was called for in which the earliest strata began accumulating more than 3 m.y. ago. This important paper by *Westgate et al.* [1990] thus demonstrates that loess in North America is potentially as significant as the Chinese deposits and will certainly help in discussions of global, as opposed to purely regional, environmental effects. It is perhaps worth emphasizing here that the basal part of the section at locality 6

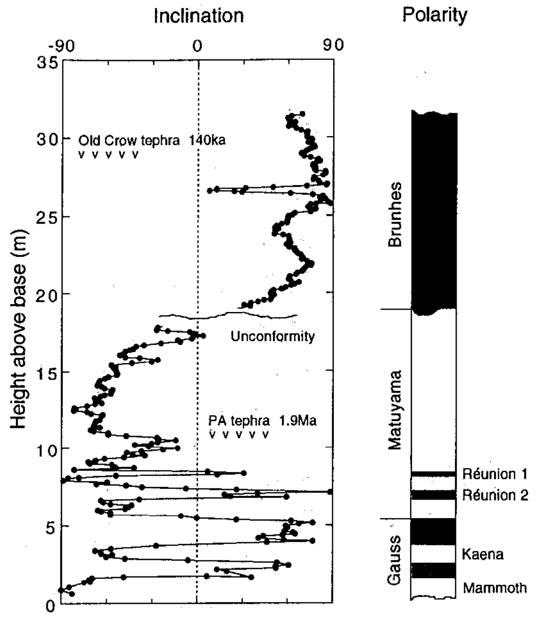

Figure 8. Magnetostratigraphy of Gold Hill loess modified after *Westgate et al.* [1990]. The inclination curve was smoothed with a three-point running average. Note the fission-track-dated tephra layers.

contains magnetozones attributed to the Kaena and Mammoth subchrons within the Gauss normal chron.

These features have also been recognized in China [*Liu et al.*, 1987; *Evans et al.*, 1991; *Zheng et al.*, 1992]. They were not discussed above, however, because we are primarily concerned with eolian loess, and in China the evidence for the detail within the Gauss comes from the Pliocene Red Clay (Poteh clay in Figure 4) underlying the loess. Loess parent material has been proposed for the Red Clay because the clay contains an increased silt fraction [*Liu et al.*, 1987] and paleosols are present with grain size distributions very similar to those of the concordantly overlying Wucheng paleosols [*Bronger and Heinkele*, 1989]. Nevertheless, a direct eolian origin has not yet been demonstrated.

In an earlier contribution, *Packer* [1979] reported results from the Craig Hills loess section in Ellensburg, central Washington State. The study was undertaken to test the feasibility of magnetostratigraphic dating as an aid to working out the poorly understood late Cenozoic geology of Washington and Oregon. Only 11 horizons were sampled over the 16 m spanned by the section, but these were adequate to delineate a reversed polarity zone in the basal 3 m, followed by normal polarity above. This zonation was interpreted as representing parts of the Matuyama and Brunhes chrons and thus provided a certain measure of useful chronological control. Before laboratory treatment, some of the reversed samples had significant normal overprints, but AF demagnetization (generally at 10 mT) successfully removed this and yielded two well-grouped antiparallel sets of magnetic vectors.

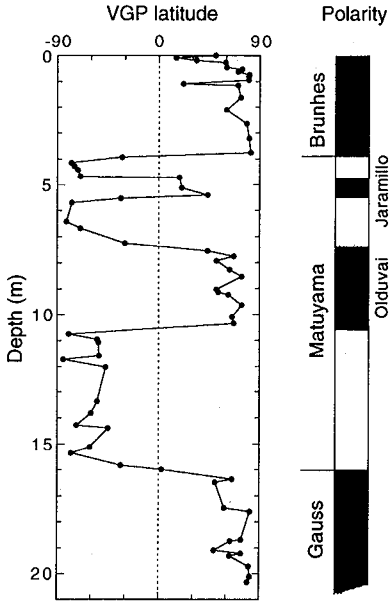

Figure 9. VGP latitudes representing a 3-m.y. magnetostratigraphic record from continental deposits including loesslike sediments on the Argentinian coast (simplified after *Ruocco* [1989]).

South America

Loess occurs extensively in Argentina, although *Pye* [1987, p. 204] claims that little is known about its composition and history. Paleomagnetic investigations have been reported by *Ruocco* [1989] and *Nabel* [1993], both of whom cite several earlier papers published in Spanish. Ruocco's work involves continental deposits exposed in coastal cliffs near Balneario Cruz del Sur (38.2°S, 57.6°W). Although the section totals only 21 m, it contains four normally polarized magnetozones and three reversed ones and can be convincingly matched to the standard polarity timescale. The Brunhes, Matuyama, and upper Gauss chrons, and the Jaramillo and Olduvai subchrons, are all clearly identifiable (Figure 9) even though accumulation rates are almost an order of magnitude lower than those found in China. The Pampean sediments are described as loesses with intervening paleosols especially in the upper part, whereas fluviatile sediments intercalate in the older parts and hiatuses cannot be excluded [*Orgeira*, 1987; *Valencio et al.*, 1987].

The conclusion that some 3 m.y. are represented is significant for the general chronological framework it provides, as indeed was the case with the early magnetostratigraphic studies in China. In the South American context this absolute chronology has been particularly fruitful as a means of dating the Pleistocene land mammal faunas, which in turn are crucial to a fuller understanding of the so-called Great American Interchange during which many species of vertebrates crossed the newly formed Panama land bridge. An excellent discussion of this important problem is given by *MacFadden et al.* [1983], who report (nonloess) magnetostratigraphic results from southern Bolivia.

Nabel [1993] focuses on the Matuyama-Brunhes boundary, which she traces some 300 km along the coast northward from the sections studied by *Ruocco* [1989], through exposures in building excavations in the cities of La Plata and Buenos Aires, to the Parana River cliffs. Of the many individual sections studied, actual paleomagnetic directions are presented for only one, the 15-m-thick sequence in Hisisa quarry in the surroundings of Baradero (33.8°S, 59.4°W). The uppermost 11 m are consistently normal with inclinations close to that expected for an axial dipole field (−53°), below which there is a rapid shift to reversed directions representing the youngest part of the Matuyama chron. As Nabel points out, the recognition of a firm reversal time line throughout Buenos Aires province has significant chronostratigraphic implications, both climatic and neotectonic; ongoing paleomagnetic work thus has considerable potential.

New Zealand

Relatively thin sequences of loess occur in New Zealand, and magnetic investigations of these have been carried out by *Pillans and Wright* [1990]. They studied a 17.4-m core collected at the southern end of the North Island about 150 km north of Wellington. The core passes through 11 distinct loess beds, each of which contains a well-defined soil in its upper portion. Since the core was not azimuthally orientated, they report inclination values only. Three dated tephra occur in the sequence, and from these they deduce an age of about 500,000 years for the base of the core. This is consistent with the normal polarities found throughout, although they had originally hoped to find the Blake event. Nevertheless, their data provide a highly coherent time series with good internal consistency, which provides a useful southern hemisphere secular variation record. Furthermore, they argue that two zones of anomalously low inclinations (in loess L_2 at about 2.5 m depth and in loess L_{11} at about 16 m depth) correspond to the Mungo and Emperor short-lived subchrons (circa 35,000 and 490,000 ka, respectively).

MAGNETIC SUSCEPTIBILITY AND PALEOCLIMATE

China

In their original paper reporting the magnetic polarity zonation of the Luochuan borehole, *Heller and Liu* [1982] presented a susceptibility profile (included in Figure 4) and briefly discussed its potential paleoclimatic significance. It was already well known that the blanket of sediments comprising the loess plateau does not consist simply of a pile of uniform loess. Excellent investigations carried out over several decades by the Institute of Geology, Chinese Academy of Sciences

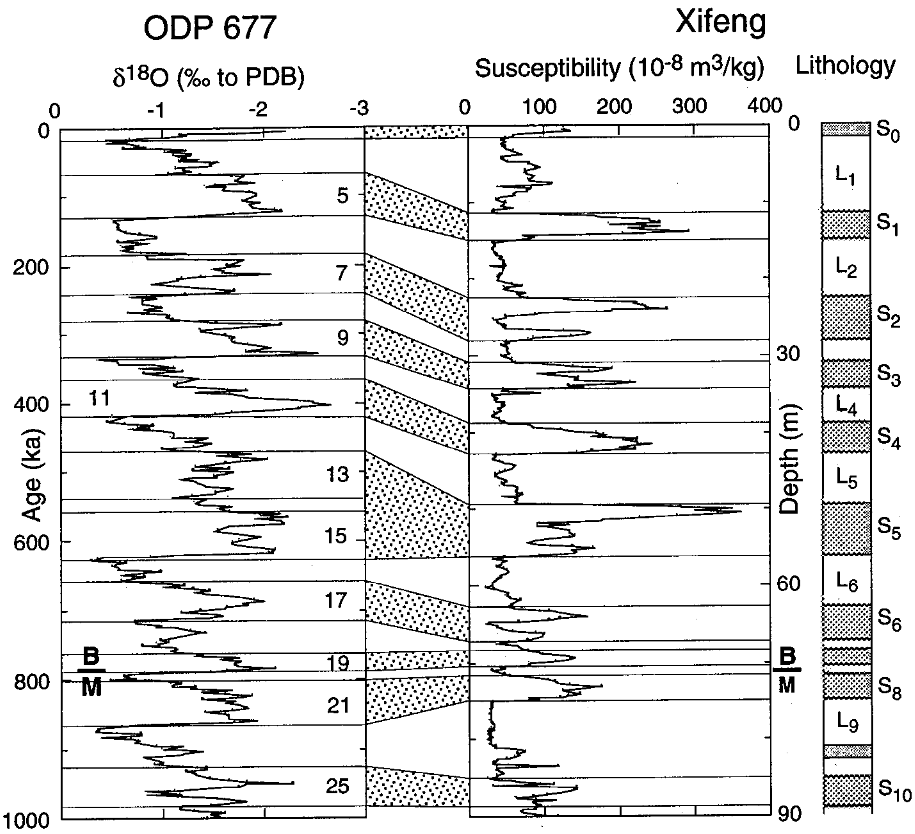

Figure 10. Oxygen isotope record at Ocean Drilling Program (ODP) site 677 (1°12′N, 83°44′W) from *Shackleton and Hall* [1989] and *Shackleton et al.* [1990] tuned to an absolute timescale using the insolation curve of *Berger* [1978] compared with the loess susceptibility profile at Xifeng (35.7°N, 107.6°E) plotted on a depth scale as measured in the field by *Liu et al.* [1988]. Note the slight offset of the M-B boundary with respect to the lithological climate indicators. The boundary is to be located in the bottom part of the warm oxygen isotope stage 19 but is observed in the cold loess layer L_8. *Forster and Heller* [1994] tried to explain this discrepancy either by a climatic phase lag between the marine and the land record or by different lock-in depths in the two sediment types eventually caused by different sedimentation rates. (PDB, Peedee belemnite standard.)

[e.g., *Liu and Chang*, 1964], had shown that the loess itself is interlayered with numerous paleosol horizons; indeed, it was the presumed climatic significance of soils that provided a major stimulus to continued loess research. Given the observed alternation of loess and paleosol strata, it was natural to enquire if these correspond to the observed highs and lows of susceptibility [*Li et al.*, 1974; *An et al.*, 1977]. It immediately emerged that there is a strong correlation, and so was initiated what has turned out to be a fruitful line of enquiry: the magnetic recording of climate.

Heller and Liu [1984] quantified the magnetic distinction between loess and soils at Luochuan by pointing out that the susceptibility of soil samples is more than twice that of loess (mean of $2.37 \pm 0.95 \times 10^{-3}$ SI units for 159 soil samples compared with $1.00 \pm 0.46 \times 10^{-3}$ for 225 loess samples). They argued that the processes leading to soil formation during relatively warm, humid intervals enhance susceptibility compared with that of pristine loess accumulated during cold, arid intervals.

The Luochuan borehole data created a great deal of interest and were duly followed up by parallel measurements in nearby gully outcrops [*Liu et al.*, 1985a; *Kukla and An*, 1989], so that very detailed coverage is now available (5- to 15-cm spacing over stratigraphic thicknesses of 150 m or more). Furthermore, comparable data are available for sections near Xifeng some 160 km to the west [*Liu et al.*, 1987; *Kukla and An*, 1989] and Baoji about 170 km to the southwest [*Rutter et al.*, 1991]. All these profiles show marked susceptibility enhancement in paleosols, and all authors agree that this represents a climatic signal, with cold, dry glacial periods when low-susceptibility loess accumulated punctuated by warm, wet intervals when high-susceptibility soils formed (Figures 4, 6, and 10).

Heller and Liu [1986] attempted a more sophisticated analysis of the borehole data by comparing them with an important independent climate proxy, the marine oxygen isotope record. For this initial test they chose the well-known record from Pacific Ocean core V28-239 [*Shackleton and Opdyke*, 1976] and claimed that each oxygen isotope stage in the marine section could be matched one-to-one with the magnetic peaks in the land section. There is indeed a distinct parallelism between these two quite different parameters: on the one hand, the isotopic composition of oxygen in $CaCO_3$ locked in the shells of tiny marine organisms; on the other, the concentration of magnetic minerals in windfall dust in China. Heller and Liu rightly conclude that the magnetic properties of Chinese loess are excellent proxy recorders of long-term environmental changes and that the Luochuan profile represents the first complete record of Quaternary climatic oscillations obtained from land.

Encouraged by this early success, *Kukla et al.* [1988] presented a detailed analysis of the Brunhes chron loess at Luochuan and Xifeng compared with the global summary of marine oxygen isotope chronostratigraphy, the so-called SPECMAP time series [*Imbrie et al.*, 1984; *Prell et al.*, 1986]. However, their analysis uses a procedure for deriving an absolute timescale which is not universally accepted. They did not use linear correlation between polarity reversals as *Heller and Liu* [1986] had done; instead, they proposed a timescale based on a particular model of loess-soil accumulation. They argue that the bulk of the susceptibility signal arises from detrital input of fine-grained magnetite from distant sources that is essentially unaffected by local conditions. Thus during warm, wet interglacial periods when dust accumulation is reduced, the susceptibility is high, whereas during glacial periods when loess accumulates more rapidly, the constant input of detrital magnetite is diluted and susceptibility is therefore reduced. The model is similar to pollution of a river downstream from a factory: when the river is in flood, the pollutants are diluted, when its flow is reduced, pollutants are more concentrated. Kukla et al.'s preferred timescale is constructed by multiplying thickness by susceptibility. This "susceptibility" timescale has been sharply criticized [*Zhou et al.*, 1990; *Maher and Thompson*, 1991] on the basis of mineral magnetic arguments (see below), so caution is necessary.

To illustrate the connection between land-based susceptibility and marine oxygen isotopes, we show here (Figure 10) profiles from Xifeng on the Chinese loess plateau and Ocean Drilling Program (ODP) site 677 from the eastern equatorial Pacific. Another excellent example is given by *Rutter et al.* [1991], who show that the Baoji loess/paleosol sequence matches very closely the composite isotope profile obtained by stacking results from four deep-sea cores (V28-238 in the western equatorial Pacific, Deep-Sea Drilling Project (DSDP) site 504 in the eastern Pacific, DSDP site 502 in the Caribbean, and DSDP site 552A in the North Atlantic). In both cases the one-to-one correlation is quite clear except for paleosol S_5, which corresponds to oxygen isotope stages 13 and 15 combined. This is in accord with the well-developed multiple character of S_5, particularly at Baoji where three distinct soil horizons are clearly visible, the upper one corresponding to stage 13 and the lower two together corresponding to stage 15. This pattern is also seen in the susceptibility variations at Xifeng (Figure 10).

Historically, one of the most striking demonstrations of the validity of the Milankovitch theory of ice ages was the recovery from marine isotope profiles of the main frequencies predicted by astronomical calculations [*Hays et al.*, 1976]. Not surprisingly, therefore, several attempts have been made at spectral analysis of susceptibility time series. Results have been mixed. *Heller and Liu* [1986, p. 1171] found that the timescale obtained by linear interpolation between polarity reversals produced a Fourier spectrum "very similar to a spectrum of a series of random numbers." In a second attempt they concentrated on the last million years (211 data points, average sampling interval of 4700 years) and modified the timescale by squeezing and stretching the time series to bring the susceptibility peaks into phase with the insolation maxima of *Berger*'s [1978] theoretical calculations based on orbital variations. This process of "tuning" is the same as that commonly applied to the oxygen isotope data [*Imbrie et al.*, 1984; *Shackleton et al.*, 1990]. Following this procedure, Heller and Liu report the appearance of a prominent peak at 40 ka, similar to that associated with the changes in obliquity (tilt of the Earth's spin axis) in standard Milankovitch theory. They do not find evidence for the precession (~20 ka) or eccentricity (~100 ka) signals. *Kukla et al.* [1990] have also attempted spectral analysis, using the maximum entropy technique [*Burg*, 1972] on the Brunhes chron portion of their stacked data. Back to about 400 ka they report well-defined periods of about 41 and 23 kyr, as expected from orbital calculations. For earlier times the precession signal is weak or absent, and the "obliquity" signal gradually drifts toward 30 ka. Given the questions about the "susceptibility" timescale, caution is again advised.

For the section at Baoji, *Wang et al.* [1990] complemented their magnetostratigraphic study [*Rutter et al.*, 1990, 1991] by determining a susceptibility profile. Fourier analysis indicates the presence of significant power near 40 and 100 ka as expected from orbital forcing. Their timescale was derived with a minimum of assumptions; two straight lines were fitted to the two obvious segments of the depth-time information obtained from the polarity reversals.

Central Asia

Very similar relations between loess lithology and susceptibility as in China are found in central Asia. Although the earlier Russian workers have noticed the susceptibility increase in paleosols [*Lazarenko et al.*, 1981], no systematic investigation has been published. The susceptibility variations in the profile at Karamaydan (Tajikistan), which have been measured in situ at intervals of 5 cm, follow the general trend observed in China [*Forster and Heller*, 1994]. The susceptibilities are about 3 times higher in the paleosols than in the loess units (paleosol average (359 samples) is 730 ± 420 $nm^3 kg^{-1}$; loess average (585 samples) is 230 ± 90 $nm^3 kg^{-1}$). Especially high values are developed in the youngest three pedocomplexes, excluding the Holocene soil. This is again similar, but not equal, to the situation on the central Chinese loess plateau, where the youngest five paleosols (excluding the Holocene soil) have markedly increased bulk susceptibility. It is also striking that the susceptibility of the unaltered loess in both regions has virtually the same magnitude. Here no time series analysis has yet been attempted, but the close correlation with the Chinese susceptibility records and the astronomically tuned oxygen isotope record at ODP site 677 [*Shackleton et al.*, 1990] permits refined dating and demonstrates the global significance of the paleoclimatic variations as recorded in the susceptibility signals.

Europe

The 55-m-thick loess sequence at Achenheim (Alsace, France) does not date back to the M-B boundary but started to accumulate some 350 kyr ago [*Rousseau and Puisségur*, 1990]. It is divided into a younger and older series. The susceptibility of the younger series, which has been dated using thermoluminescence and ^{14}C methods and covers the last climatic cycle (i.e., the last 128 kyr) has been measured by *Rousseau et al.* [1994] with a portable susceptibility meter. Although the susceptibility variations are not as strongly developed (maximum susceptibility is 110 $nm^3 kg^{-1}$), they are similar to those observed in central China. Comparison with the oxygen isotope scale shows that the loess record covers the isotope stages 2–4 completely, whereas the equivalents of some substages of stage 5 are missing owing to hiatuses in the sequence.

Susceptibility profiles in the Polish loess/paleosol sequences near Lublin (SE Poland) show reversed relations of susceptibility enhancement as compared with China [*Maruszczak and Nawrocki*, 1991; *Nawrocki*, 1992]. Volume susceptibilities never exceed values of 0.4×10^{-3} SI units and are, in general, higher in the nonweathered layers. They are found drastically diminished in the upper horizons of paleosols. The Polish soils have accumulated in a more humid and cooler climate of the periglacial realm as compared with the drier and warmer peridesertic climate of China or central Asia. This climate favored gleyification and leaching of the soils during interstadial and interglacial times, leading to degradation and destruction of the magnetic carriers.

North America

The first unambiguous evidence that magnetic susceptibility of land-based stratigraphic sequences provides a climate proxy that can be directly linked to Milankovitch cyclicity was reported by *Begét and Hawkins* [1989] from the important Halfway House exposure near Fairbanks, Alaska. They showed that spectral analysis of the 16-m susceptibility profile yields the main Milankovitch periodicities, even though the profile probably represents no more than ~250 kyr. One very important difference between the Alaskan and Chinese data must be noted: whereas the paleosols in China invariably have higher susceptibility values than the loess, the opposite relationship holds in Alaska. In a more detailed paper, *Begét et al.* [1990] attributed this to variable wind action. They argue that on average, winds were stronger during glacial intervals and were thus able to transport more of the larger magnetite grains from the Tanana River flood plain, which is the source of the loess found at Halfway House. It is the coarser magnetite that is apparently responsible for the higher susceptibility in Alaskan loess; this was demonstrated by the observed correlations between susceptibility and loess grain size (at the ninetieth percentile, correlation coefficient R = 0.78) and between susceptibility and the average size of the five largest magnetite grains separated from the loess (R = 0.92). Basically, their argument is that during glacial periods winds are stronger, and hence more coarse magnetite grains are entrained. When they are eventually deposited in loess, these grains give rise to higher susceptibility. On the other hand, during interglacials (when soils form), winds are weaker and so therefore is susceptibility. Stronger winds in Alaska during glaciations are consistent with numerical models of global climate [*Kutzbach*, 1987] and are supported by geological evidence [*Thorson and Bender*, 1985].

Atmospheric activity also plays the key role in the susceptibility variations observed in late Wisconsin (25–14 ka) loess sequences of Indiana investigated by *Hayward and Lowell* [1993], which do not seem to be influenced by pedogenic alteration. Susceptibility lows (typically 45 to 55 $\times 10^{-5}$ SI units) characterize incipient paleosols that record periods of low loess accumulation, whereas moderate to high susceptibilities (typically 75 to 110 $\times 10^{-5}$ SI units) correspond to stratified or massive loess reflecting periods of intermediate to high loess accumulation rates. Hayward and Lowell attribute the variable detrital magnetite enrichment and loess accumulation to changing wind intensity and soil moisture. They give preference to the influence of rainfall, since no systematic variation of magnetite grain sizes has been noticed.

The "wind variability" model is applicable in Alaska and in locations near ice sheets, such as those in Indiana, but may not be appropriate for other areas. For example, large detrital magnetite grains do not seem to be important in the central Chinese loess plateau, where pedogenetic processes control the loess magnetic properties. This topic has received a great deal of scrutiny, and we therefore return to discuss it at length in a later section.

The susceptibility-climate relationship already changes in the relatively mild and moist paleoclimate of the Peoria loess (25–14 ka) of western Nebraska at Eustis [*Rousseau and Kukla*, 1994]. Detailed investigations of mollusc assemblages demonstrate that the paleoclimate changed during deposition of the 20-m-thick loess from moderately cool and moist to extremely cold and dry. This is reflected as a gradual bulk susceptibility decrease by about a factor of 2. Rousseau and Kukla attribute this susceptibility reduction to a gradual increase of the initially low dust deposition combined with increasing humidity, which leads to partial destruction of magnetite when the loess becomes gleyed, or reduced in stagnant water.

South America

In addition to the magnetic zonation of the coastal cliffs in Argentina, *Ruocco* [1989] also reports a single 17-m susceptibility profile. From her diagram it appears that low values are preferentially associated with paleosol horizons, although there are only four soils involved. This tentative observation merits further scrutiny, as it may be another example of control by wind variability as proposed by *Begét et al.* [1990] for interior Alaska.

New Zealand

Pillans and Wright [1990, p. 182] measured susceptibility of samples taken from the 17.4-m drill core used in their magnetostratigraphic investigation and report "high values in paleosols and low values in loess." However, they ascribe this primarily to volcanic input from several nearby sources.

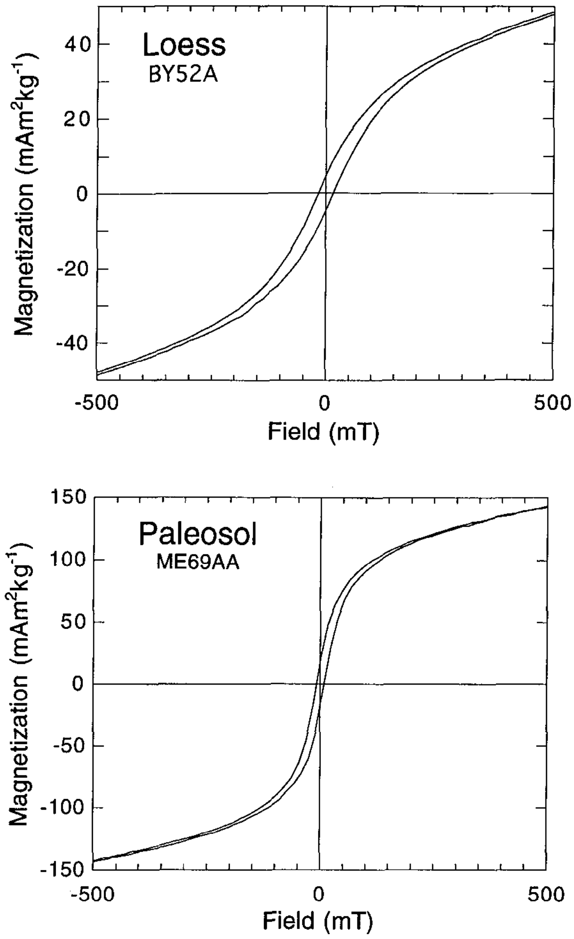

Figure 11. Examples of hysteresis loops for a fresh and unaltered loess sample from Baicaoyuan on the arid western Chinese loess plateau and a strongly altered paleosol sample from Baoji in the moist south of the plateau, as measured with an alternating gradient magnetometer. Both curves are uncorrected for paramagnetic contributions and have been measured up to ±2000 mT. Note that the magnetization scales differ by a factor of 3. The saturation magnetization M_s obtained after subtraction of the paramagnetic contribution is 34 mA m^2 kg^{-1} for the loess sample and 131 mA m^2 kg^{-1} for the paleosol.

MINERAL MAGNETISM

In view of the important geomagnetic and paleoclimatic information provided by loess deposits, it is necessary to understand the physical basis of these signals. In which components of the sediments is the information stored, and what is their origin? Both aspects, geomagnetic and paleoclimatic, remain the subject of active debate, and there can be little doubt that the final word has not yet been heard in either case.

First, consider the source of the climatic susceptibility signal. As pointed out above, *Heller and Liu*'s [1982] original paper recognized that paleosols have susceptibility values about twice those of unmodified loess. Since paleosols formed under conditions that can be generally characterized as warm and wet, whereas unmodified loess represents a dry, cold environment, the concept of susceptibility as a climate proxy was born. Indeed, such a concept could have been expected from earlier work in soil science, which had demonstrated enhanced susceptibilities in topsoil compared with the underlying subsoil and parent bedrock [*Le Borgne*, 1955; *Mullins*, 1977]. The contrast between soils and loess is further demonstrated by magnetic hysteresis properties (Figure 11). As pristine loess is transformed to soil, we see a marked decrease from a coercive force H_c of 18.4 mT to a value of 8.5 mT and a corresponding increase in remanent saturation magnetization M_{rs} from 4.6 mA m^2 kg^{-1} to 18.6 mA m^2 kg^{-1}; production of soils leads to stronger, but softer, magnetization. The increasing saturation magnetization signifies a fourfold growth of the amount of magnetic low-coercivity material.

All authors that have worked on Chinese loess are agreed on the empirical basis of the magnetic climate proxy; the question is, what is the underlying cause?

Early reports [*Heller and Liu*, 1982, 1984] appealed to the relative enrichment of magnetic particles by the leaching of carbonates from the paleosols. This may be a minor contributing factor but cannot possibly lead to the observed susceptibility increases by factors of 2 or more because initial carbonate content is never more than about 20%. A second suggestion supposes that the magnetic content is supplied at a constant rate from a remote, but unspecified, source. In this model, susceptibility variations are due to fluctuations in the relative amount of dust particles arriving from the immediate hinterland. During cold, dry intervals, large quantities of loess are blown in. When it is warm and wet, vegetation cover and moisture inhibit the dust flux. Paleosols therefore have higher susceptibilities, whereas unmodified loess dilutes the magnetic content and leads to lower values. *Kukla et al.* [1988] first proposed this concept and went on to put forward the idea of a "susceptibility timescale" derived by multiplying the thickness of each layer by its susceptibility. While there is some merit in the idea of reduced dust flux during warmer periods, this model cannot account for all the observations.

As was originally pointed out by *Zhou et al.* [1990], paleosols are characterized by a higher content of small magnetic grains than the intervening loess beds. Variations in laboratory remanences and frequency dependence of susceptibility indicated that the different magnetic components present in loess and paleosols occur in different relative proportions, a situation that cannot arise from simple concentration or dilution. *Zhou et al.* [1990] therefore drew the important conclusion that magnetic enhancement in paleosols reflects (in part, at least) some form of in situ process taking place during soil formation. It is now generally agreed that magnetic changes due to pedogenesis are crucial to a proper understanding of magnetic climate proxy data [*Heller et al.*, 1991; *Liu et al.*, 1990, 1991a, b; *Maher and Thompson*, 1991].

Maher and Thompson [1992] identify magnetite (Fe_3O_4) as the main pedogenic mineral responsible for susceptibility enhancement. This conclusion is based on isothermal and anhysteretic remanence acquisition in addition to susceptibility data; their procedure employs a specific multiple regression technique to fit the experimental information to corresponding data for synthetic analogues [*Thompson*, 1986]. For four loess strata (11 samples) and four paleosols (10 samples), they find about twice as much magnetite in soils as in loess (mass fractions of $0.11 \pm 0.01\%$ and $0.053 \pm 0.007\%$, respectively). Furthermore, they go on to argue from an analysis of variance that over 90% of the susceptibility contrast between soil and loess is due to magnetite grains lying near the superparamagnetic threshold (~30 nm for equidimensional grains). Such grains, which are not able to keep information about the paleomagnetic field, are well below the optical limit, but *Maher and Thompson* [1992] demonstrate their presence by means of electron microscopy. We have also followed this approach, with typical results as illustrated in Figure 12. *Hus and Bai* [1993] report data which confirm the pedogenic production of magnetite; in particular, by studying size fractions of disaggregated samples, they demonstrate the importance of smaller grains. Unfortunately, the smallest grains that can be effectively separated using standard methods have dimensions up to ~2 μm, which exceeds by almost 2 orders of magnitude the size range considered important by *Zhou et al.* [1990] and *Maher and Thompson* [1992]. *Banerjee et al.* [1993] confirm the observation that bulk susceptibility is more frequency-dependent in paleosols S_1 and S_0 than in loess layers L_2 and L_1, again indicating the importance of ultrafine particles. They also explore the superparamagnetic region below ~30 nm by cooling samples down to 15 K and monitoring the loss of saturation remanence during rewarming to room temperature. In this way, they directly demonstrate the greater importance of superparamagnetic grains in paleosol layers compared with their parent loess beds; the temperature dependence of magnetization observed also shows the sharp drop near 120 K due to the Verwey transition, demonstrating that the material in question is undoubtedly Fe_3O_4.

In their early work on the Chinese loess, *Heller and Liu* [1984] recognized that a magnetically "hard" fraction is present with coercivities well beyond those typical of magnetite. Thermal demagnetization yields maximum unblocking temperatures above 600°C, indicating that hematite (α-Fe_2O_3) is present, which is confirmed by optical microscopy. The analysis of *Maher and Thompson* [1992] also identifies hematite in significant quantities but suggests little difference between loess and paleosol (mass fractions of 0.92% and 0.73%, respectively). These amounts exceed the corresponding values for magnetite, but it must be remembered that magnetite is 250 times more magnetic than hematite. The natural and laboratory remanences certainly have a hematite contribution, but the susceptibility enhancement in paleosols is due to a magnetically "soft" mineral, namely, magnetite and/or maghemite (γ-Fe_2O_3). This follows from the work already cited [*Zhou et al.*, 1990; *Maher and Thompson*, 1992; *Banerjee et al.*, 1993] and is further supported by *Liu et al.* [1993], who report detailed rock magnetic characteristics of samples from the uppermost 16 m of the section at Xifeng spanning the L_2-S_0 interval. In addition to magnetite and hematite, Liu et al. argue for the presence of minor amounts of maghemite. This mineral was not detected by *Maher and Thompson* [1992] because their multiple regression procedure did not include appropriate parameters for synthetic maghemite samples; their model simply excludes this mineral from the outset. However, it is interesting to note that they obtain an average mass fraction for magnetite of 0.053%, close to the combined amounts

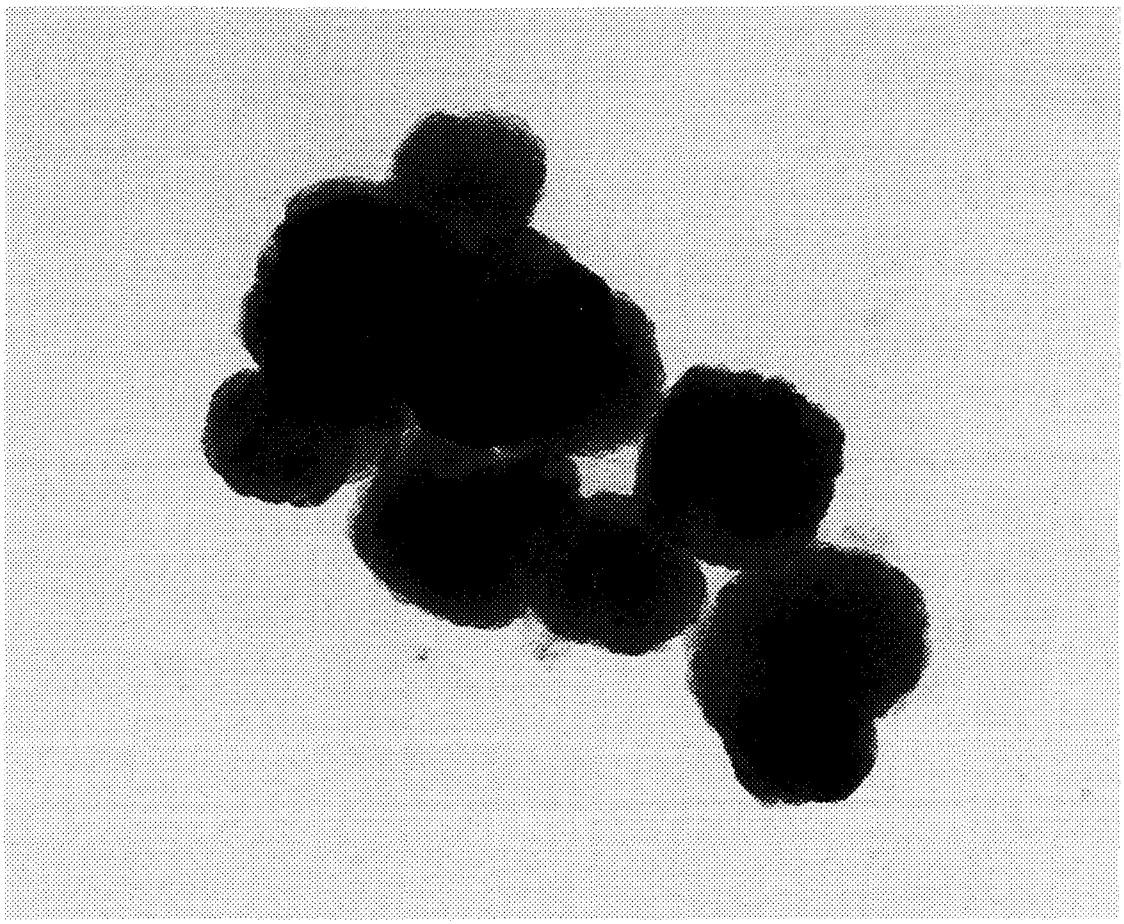

Figure 12. Electronmicrograph of magnetite crystals separated from pedocomplex 2 at Karamaidan. Linear magnification is ×129,000 (i.e., diameter of largest crystal is ~230 nm).

of magnetite and maghemite (0.046%) reported by *Evans and Heller* [1994].

The Mössbauer spectroscopic studies on magnetic extracts by *Vandenberghe et al.* [1992] also show predominantly hematite and magnetite as the main iron-bearing minerals in a section at Jixian (central Chinese loess plateau), but they have also identified substantial amounts of maghemite in paleosol S_1. *Verosub et al.* [1993] have extended the idea of a maghemite contribution by considering the possible significance of this mineral as the basis of the climate proxy (enhanced susceptibility) signal, not simply as a minor constituent of unaltered parent loess. To seek support for this suggestion, they subjected about 30 Chinese loess/paleosol samples to leaching with citrate-bicarbonate-dithionite (CBD). This is a soil chemistry procedure developed to extract iron oxides from clay samples prior to X ray analysis. It provides semiquantitative analyses [*Borggaard*, 1988]. Susceptibilities decrease by 59–83% after treatment and yield an almost constant value of 0.15 $\mu m^3\ kg^{-1}$ throughout the 140-m Luochuan section studied. This CBD-insoluble material is interpreted as an original air fall input which has not been affected by pedogenesis. The identification of the CBD-soluble material, however, which is increased by up to a factor of 2 in the paleosols and accounts for about one third of the total iron in loess (Figure 13), remains uncertain, since both maghemite and ultrafine magnetite grains are removed by the procedure. Thus, although the soil chemistry approach strongly supports the pedogenic model of susceptibility enhancement, it has not, so far, succeeded in positively identifying the mineral(s) responsible. As *Verosub et al.* [1993, p. 1014] conclude, "additional work is needed in this regard."

Despite these difficulties, *Eyre and Shaw* [1994] renew the argument for maghemite as the source of enhanced susceptibility in paleosols as compared with parent loess material. Their motivation is a desire to explain what they describe as "anomalous" thermomagnetic behavior, rather than any compelling evidence for the direct identification of this mineral. The problem is that observed heating and cooling curves for susceptibility-enhanced material are reversible, indicating high thermal stability, whereas ultrafine magnetite ought to oxidize to maghemite upon heating [*Özdemir et al.*, 1993]. Furthermore, there are reports that maghemite can be thermally stable if the grain

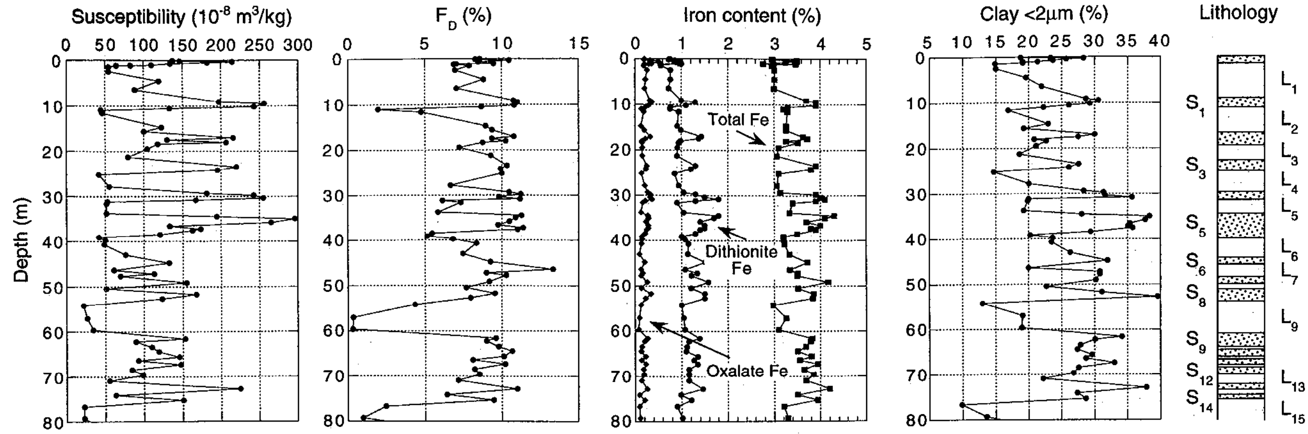

Figure 13. Susceptibility, frequency dependence of susceptibility (F_D), iron content (presented as total iron, dithionite, and oxalate-soluble iron), clay content, and lithology of alternating paleosol (S) and loess (L) beds as a function of profile depth at Luochuan. Nonmagnetic data are from *Heinkele* [1990]. Total depth corresponds to about 1 Ma. Note the close correlation between susceptibility and F_D but also between the magnetic parameters and the chemical, mineralogical, and lithological properties. Oxalate-soluble iron documents the content in amorphous iron hydroxides and oxides, whereas dithionite soluble iron is due to pedogenic iron oxides.

sizes fall in the range 10–30 nm [*Ayyub et al.*, 1988]. A serious difficulty arises, however, when it is realized that the thermomagnetic curves themselves offer strong support for the predominance of magnetite. For example, *Evans and Heller* [1994] studied the L_4/S_3 couplet at three sites spread across the loess plateau and show that a representative sequence of thermomagnetic curves (both reversible and irreversible types) all yield Curie points (determined objectively by the method of *Moskowitz* [1981]) within 10°C of the value for pure magnetite. Clearly, the magnetite-maghemite debate is not yet settled, although convergence is taking place in other aspects of mineral magnetism.

Like *Eyre and Shaw* [1994], *Evans and Heller* [1994] interpret their results of acquisition of IRM in terms of two distinct ferrimagnetic components, A and B. Component A is the original air fall contribution, amounting to a mass fraction of slightly less than 1% and consisting of hematite (0.88%), maghemite (0.024%), and magnetite (0.022%). This component appears to be essentially uniform across the entire Chinese loess plateau and even into central Asia [*Forster and Heller*, 1994]. Component B comprises ultrafine (below ~100 nm) magnetite particles with remarkably homogeneous characteristics but varying in amount by an order of magnitude from the northwest to the southeast (Figure 14), a distance of ~350 km. Comparison with appropriate hysteresis, susceptibility, and viscous decay data for synthetic magnetite grains [*Maher*, 1988; *Dunlop*, 1983] implies that a mass fraction of about 0.5% is all that is required to account for the most magnetic site. Furthermore, the remanence properties are very similar to those reported by *Petersen et al.* [1986] for Fe_3O_4 particles produced by magnetotactic bacteria. This being so, observing the amazingly uniform magnetic properties of the ultrafine magnetite fraction, *Evans and Heller* [1994] argue that the magnetic enhancement resulting during the production of S_3 from its parent loess, L_4, is bacterial in origin and reflects the concomitant climatic shift to warmer and wetter conditions. Direct evidence of bacterial magnetosomes, however, as found in modern soils [*Fassbinder et al.*, 1990], has hardly been demonstrated for loess strata [*Maher et al.*, 1994b]. It is questionable anyway whether large enough quantities of chains of bacterial magnetite crystals would be preserved in their organic sheaths for microscopic observation and would not be mostly destroyed in the highly bioturbated paleosols.

Regardless of the actual mechanism underlying the magnetic climate proxy, it has recently been observed that the frequency dependence of susceptibility is remarkably uniform, not only throughout the Chinese loess but also at loess sites in central Asia and even in soil chronosequences along the California coast [*Forster et al.*, 1994]. A close analysis of the decrease in measured susceptibility with increasing frequency of the field used (in most cases only the two frequencies of the commercially available Bartington susceptibility meter are employed, ~0.5 kHz and ~5 kHz) indicates that the conventional index F_D (= $(\chi_{LF} - \chi_{HF})/\chi_{LF}$) × 100%, where χ_{LF} and χ_{HF} are the low-frequency and high-frequency susceptibilities, respectively) is dependent on the concentration of superparamagnetic grains (Figure 13) and is somewhat misleading. When allowance is made for a small frequency-independent fraction, all sites yield very similar true frequency dependence (10.9 ± 1.0%). The full meaning of this uniformity, which points to a uniform grain size dis-

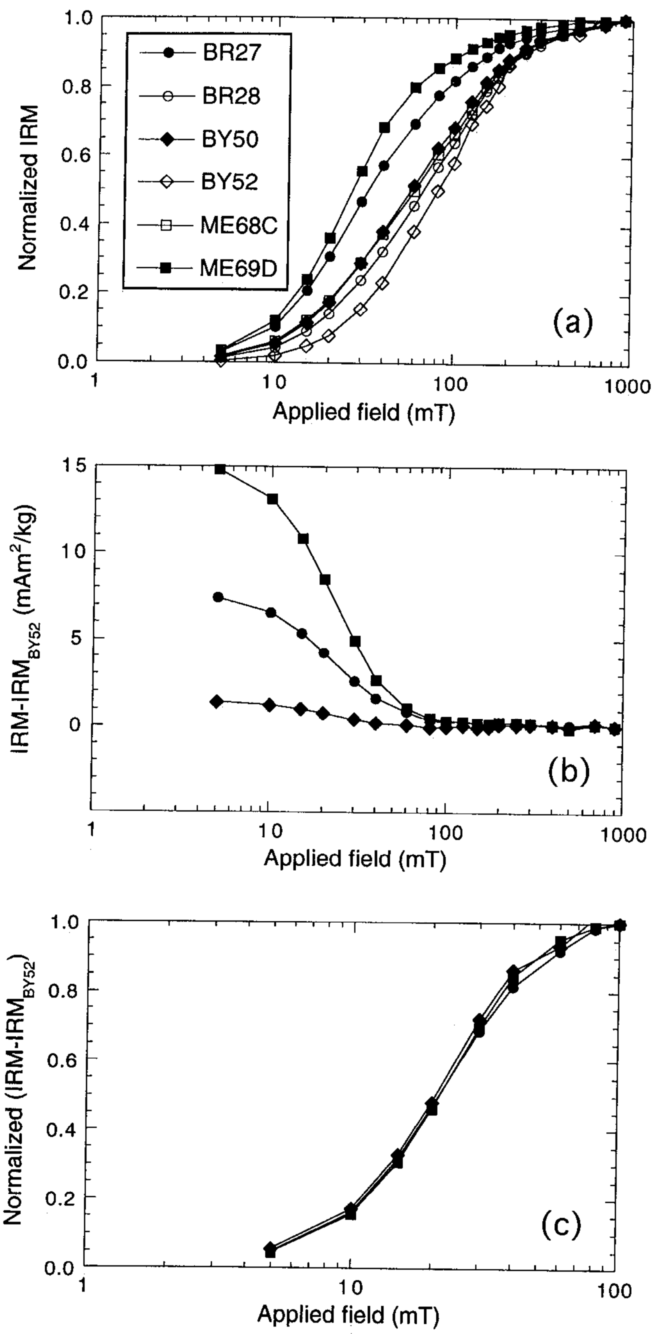

Figure 14. Isothermal remanent magnetization (IRM) acquisition of samples taken from the loess/paleosol couplet L_4/S_3 at three different outcrops on the loess plateau. Symbols are diamonds, Baicaoyuan (BY); circles, Luochuan (BR); squares, Baoji (ME). Solid symbols indicate S_3; open symbols indicate L_4. (a) Normalized cumulative curves for three paleosol and three loess samples. (b) The pedogenic component B in the paleosol samples is obtained by subtracting the uniform air fall IRM component A. (c) Normalized cumulative curves for the pedogenic component B in the three paleosol samples.

tribution [*Stephenson*, 1971a, b], is not yet entirely understood and is still under investigation.

In contrast to the vigorous enquiries into the nature of the paleoclimatic signal derived from magnetic susceptibility measurements, the corresponding situation concerning the magnetostratigraphic signal derived from remanence measurements is much less advanced. One problem is that in China, where the bulk of the data currently available comes from, there is widespread magnetic overprinting. This means that upon initial measurement of the natural remanent magnetization (NRM) in the laboratory, very few samples yield reversed polarity; everything has been overwhelmed by the secondary, normal-polarity component acquired during Brunhes time (Figure 15). The capacity of loess and paleosol material to acquire viscous remanence was documented by *Heller and Liu* [1984], who showed that viscosity coefficients measured in the laboratory suffice to account quantitatively for the strong Brunhes overprint observed. The overprint opposes the original remanence of reversely magnetized samples and causes them to have systematically weaker NRM values. A good example is provided by more than 400 samples from the Baoji section [*Rutter et al.*, 1991]. Median values for the NRMs of normal and reversed samples differ by a factor of 2.8, indicating (by simple vector subtraction) that the overprint is approximately twice as large as the original remanence. *Forster and Heller* [1994] arrive at similar conclusions for the Karamaydan section in the Tajik depression.

Despite the heavy overprints, it has been possible to uncover the correct magnetostratigraphy by means of incremental demagnetization in the laboratory. For the Chinese samples this always requires thermal demagnetization (usually at ~300°C), with very few samples responding to alternating field treatment (Figure 16). In Tajikistan, on the other hand, AF cleaning was found to be effective [*Forster and Heller*, 1994]. The failure of AF cleaning in China led *Heller and Liu* [1984] to suggest that the original remanence is a chemical remanent magnetization (CRM) residing in hematite. They point out that if the process of hematite growth is sufficiently slow, short-lived polarity subchrons such as the Blake may easily be obscured. This was in accord with earlier studies that failed to find evidence of such features, although tremendous effort employing closely spaced sampling across critical profiles had been devoted to the problem. On the basis of lateral variations in the stratigraphic position of the M-B boundary, *Heller et al.* [1987] argue that the remanence lock-in time is of the order of 10^4 years. However, *Zhu et al.* [1994a] have recently presented a convincing record of the Blake short-lived polarity subchron at Xining, western China. Reasonable estimates of the local deposition rate suggest that the Blake lasted 5300 ± 600 years, so that the remanence acquisition process, at least for this site, was relatively rapid; indeed, detail within the subchron itself suggests that features as short as a few centuries can be recovered.

Clearly, there are differences between sites. These

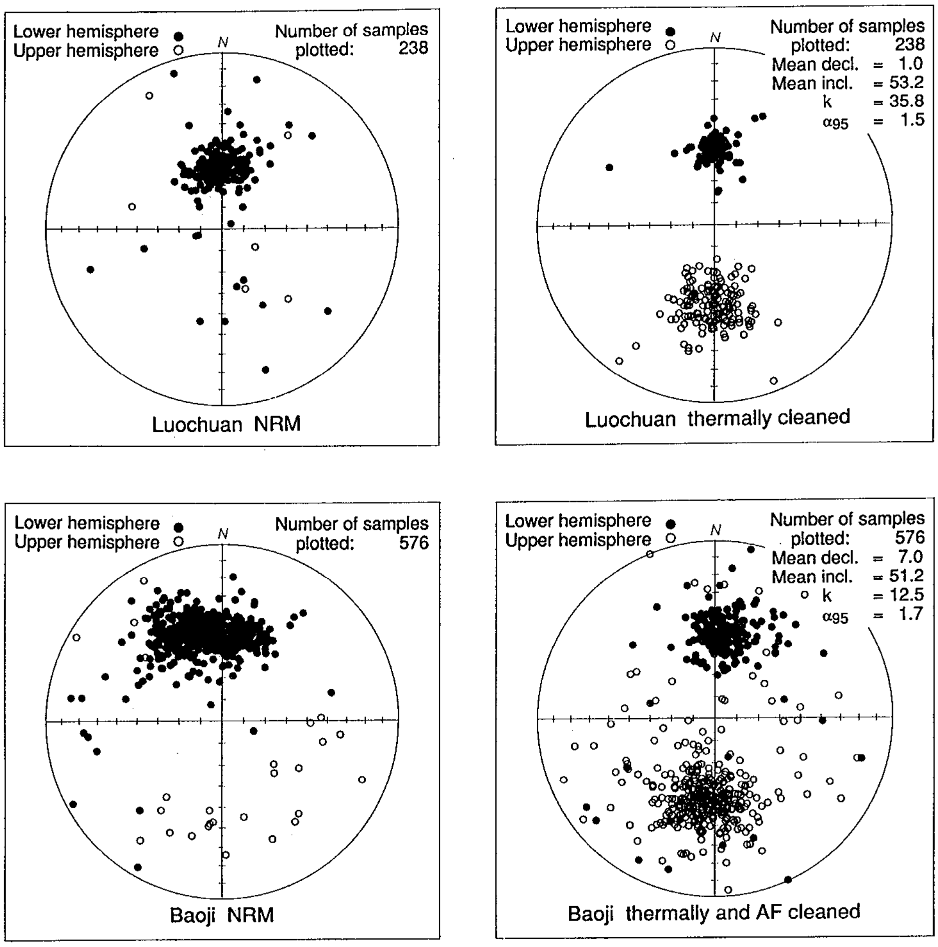

Figure 15. Stereographic projection of NRM directions before and after cleaning as recorded at Luochuan and Baoji. The uncleaned NRM is dominated by a strong present-day overprint at both sites. The cleaned reversed directions are more scattered than the normal directions, probably because of incomplete removal of the secondary component. Thermal cleaning at Luochuan results in reduced directional dispersion as compared to the Baoji data obtained by using thermal and AF demagnetization. Here k and α_{95} represent *Fisher*'s [1953] statistical dispersion parameter and 95% confidence angle for the mean direction, respectively.

can probably be attributed to variable local conditions. For example, Luochuan, where no short subchrons are found, lies in an area that currently receives an annual rainfall of 620 mm, whereas Xining lies in the far western arid region (350 mm). Possibly, the generally more moist and warmer climate in the central plateau promotes oxidation and creates partly oxidized phases that are strongly resistant to AF demagnetization. Such a process, which has led to increased coercivity of secondary magnetizations, has been invoked for the partially remagnetized marine Trubi marls in Sicily [*Van Velzen and Zijderveld*, 1992]. It is very likely that weathering, leading to chemical changes such as oxidation of magnetite [cf. *Vandenberghe et al.*, 1992], is subdued in the west where depositional remanent magnetization (DRM) might have been the prominent magnetization process and therefore higher magnetostratigraphic resolution is retained. Resedimentation of Chinese loess under wetting conditions in the laboratory produces, in addition to a viscous component, a stable DRM as soon as the sediment dries out [*McIntosh*, 1993]. Such a postdepositional magnetization may migrate upward in the sediment pile and create a coherent record of the geomagnetic field, the temporal resolution of which is dependent on the accumulation rate. It is fair to say, however, as *Zhu et al.* [1994a, p. 700] conclude, that "more research is needed to understand the exact processes of remanence acquisition of loess deposits."

DISCUSSION

Magnetostratigraphy and Deposition History

Given that continuous sedimentation records are available, magnetostratigraphy is one of the few reliable dating tools in Quaternary geology, especially when sediment ages exceeding 100 kyr prohibit accu-

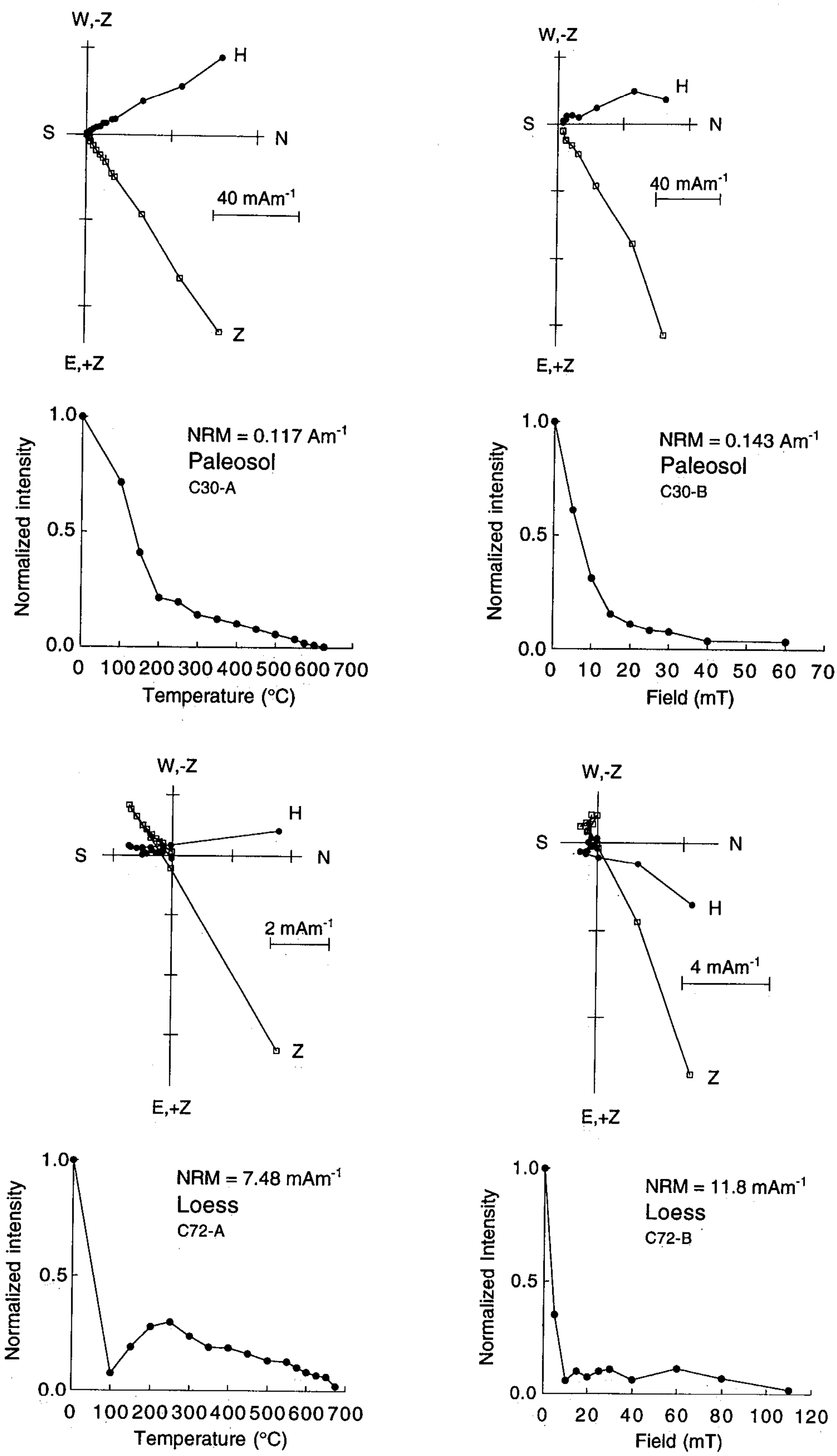

Figure 16. Thermal and alternating field demagnetization of loess and paleosol samples from Luochuan presented as orthogonal vector plots and intensity decay curves. A strong secondary component of normal polarity which has low coercivity (~10 mT) and low maximum unblocking temperature (~200°–250°C) is observed in all samples. Maximum unblocking temperatures of the characteristic NRM component significantly exceed 575°C, indicating a hematite contribution to the NRM (compare the reversed sample C72-A). AF-cleaned samples at Luochuan rarely exhibit well-defined characteristic NRM directions.

rate absolute age dating using thermoluminescence methods [*Wintle*, 1990]. Indeed, stratigraphic age assignment based on magnetic polarity has completely revised our ideas concerning the timing of loess formation.

Estimates of the beginning of Chinese loess deposition have more than doubled (from 1.2 Ma to 2.6 Ma) since the early 1980s, when the loess magnetozone pattern was first successfully compared to the geomagnetic timescale (Figure 3). The increased age is compatible with the onset of ice-rafted detritus in the oceans due to initiation of major glaciation in northern Europe [*Backman*, 1983]. The M-B boundary is observed in many Chinese loess sections in the same lithological or paleoclimatological unit L_8 (sometimes S_8), confirming the homogeneous sedimentation structure of the Chinese loess plateau. The slight stratigraphic discrepancy between L_8 and S_8 [cf. *Zheng et al.*, 1992] may readily be explained by a variable NRM lock-in depth. The Jaramillo subchron is also confined to specific loess beds including the paleosols S_{10} and S_{11}. In central Asia, loess sedimentation seems to have commenced at a later stage just prior to the Olduvai subchron, or even just prior to the Jaramillo subchron as in Uzbekistan.

Using the magnetostratigraphic ages, high loess accumulation rates are derived for the huge sediment piles in China and central Asia. The most important sections in the central loess plateau yield a Brunhes average accumulation rate of $\sim 7.5 \pm 1.3$ cm kyr^{-1}; in the Matuyama, $\sim 4.9 \pm 0.6$ cm kyr^{-1} is the measured value. In the western loess plateau, even higher rates are recorded: ~ 26 cm kyr^{-1} during the Brunhes chron and 8.2 cm kyr^{-1} for most of the Matuyama chron according to *Rolph et al.* [1989], or 33 cm kyr^{-1} according to *Burbank and Li* [1985]. The Brunhes accumulation rate in Tajikistan is of the order of 10.5 cm kyr^{-1}, and during the Matuyama, 7.4 cm kyr^{-1} is observed. The trend toward increased loess sedimentation occurs at about the same time when the amplitudes of the marine oxygen isotope and the susceptibility signals imply stronger climatic fluctuations. The land record may be related to changing monsoonal wind conditions associated with the uplift of the Tibetan plateau and neighboring mountain ranges in the late Quaternary [*Pye*, 1987].

Similar ages have been reported for the European loess, which has been studied systematically since the 1970s. The Olduvai subchron is apparently the key feature for the beginning of loess sedimentation in Austria and the Czech Republic, although the abrupt sedimentation change at the M-B boundary emphasized above casts some doubt on the correlation of the subchrons and necessitates a sedimentation break. The assumption of a continuous and essentially constant sedimentation, however, with the Brunhes accumulation rate of about 8 cm kyr^{-1} would set the Jaramillo subchron at a depth of about 75 m at Krems, where previous interpretations have placed the Olduvai. This basal age of approximately 1 Ma rather nicely complements the oldest ages of the nearby Hungarian loess, which start immediately after the Jaramillo subchron.

The very young midcontinent loess deposits in North America contrast with the very old Alaskan deposits, which date back far into the Gauss chron, as do some of the Argentinian loess columns. Thus magnetostratigraphic evidence suggests that on a global scale there are enormous differences in the onset of loess deposition during the Quaternary for which wind distribution patterns, tectonic activity, and simply geographic position with respect to the source areas must play major key roles.

Geomagnetism

As far as the geomagnetic field itself is concerned, the key observations emerging from loess studies are the superb polarity zonations stretching back more than 2.5 m.y. into the Gauss chron. These have been thoroughly dealt with above; here we turn to other related aspects. For 238 samples at Luochuan, *Heller and Liu* [1984] show that after thermal demagnetization the two polarity groups are almost antiparallel (Figure 15, Luochuan, normal polarity declination $D = 359.7°$, inclination $I = 55.6°$, $\alpha_{95} = 2.2°$, $n = 88$; reversed polarity $D = 181.7°$, $I = -51.8°$, $\alpha_{95} = 2.1°$, $n = 150$) and close to the direction of a geocentric axial dipole ($D = 000°$, $I = 55.3°$). In the southern hemisphere, *Pillans and Wright* [1990] similarly report that 450 samples from their 17.4-m core are all of normal polarity and have a mean inclination ($-61°$) not significantly different from that of an axial dipole ($-59°$).

In addition to the major magnetozones, evidence has also been sought for the many shorter subchrons which various authors have reported. Here the loess record is not particularly helpful. The most convincing example is the recently reported record of the Blake subchron [*Zhu et al.*, 1994a] as discussed above in connection with the magnetization process responsible for loess remanence. Over a stratigraphic thickness of 55 cm in the Xining section a complex pattern is observed, consisting of three clearly defined periods of reverse polarity separated by two short intervals of normal polarity. Details of this kind (not always identical) have been reported from several locations around the world, so the Xining loess data represent an important record of the Blake. Other short-lived subchrons have not been so convincingly identified in the loess record. *Rolph et al.* [1989] and *Zheng et al.* [1992] report appropriate data from sections near Lanzhou and Xian, respectively. Their interpretations collectively include many of the polarity features listed by *Champion et al.* [1988], but the evidence is not compelling. Most cases rely on single samples. More detailed sampling is necessary to demonstrate the reality of sustained polarity intervals interpretable as valid geomagnetic signals. A further problem is the possibility of sedimentary disturbances (ice wedging, involutions, creep, or solifluction), which can generate divergent remanence vectors easily mistaken for geomagnetic signals [*Hus et al.*, 1993].

Quasi-continuous, and relatively rapid, accumulation makes the Chinese loess a promising target for detailed investigation of the actual transition process from one polarity state to the other. Accordingly, *Sun et al.* [1993] report results from the M-B boundary in a section at Xifeng. The transition spans 380 cm, and the estimated accumulation rate indicates that some 20 kyr is represented. The directions within the transition exhibit a complex pattern in which the polarity switches back and forth five times. If this can be confirmed by corresponding results from other sections, it potentially represents an observation of the greatest importance. At the present time, however, caution is advisable, since the possibility of inadequate removal of magnetic overprints offers just as likely an explanation, particularly when it is noted that the apparent intermediate directions are mostly artifacts of the five-point smoothing filter employed.

A more convincing example involving the termination of the Olduvai subchron is reported by *Liu et al.* [1993] from the Wangning section (37.1°N, 112.7°E). Over a stratigraphic distance of ~60 cm a sequence of genuine intermediate directions indicates that the virtual geomagnetic pole first moved back and forth from the north pole approximately along the meridian of Moscow with an amplitude of about 60°. It then jumped abruptly to the central Pacific (based on only two horizons), before lingering for some 10 kyr (six horizons spanning 30 cm) in a steady position just south of Madagascar. This is reminiscent of the findings of *Hoffman* [1992], although the regions he observes as stable intermediate pole positions are situated near South America and Australia.

At Lanzhou, *Rolph* [1993] describes a record of the initiation of the Jaramillo subchron. The resulting declination and inclination magnetograms are rather noisy with many rapid fluctuations, often represented by single data points. Nevertheless, a reasonably coherent path from reversed to normal polarity takes place over a distance of about 1 m, but he concludes that on the question of preferred transitional VGP paths the evidence is inconclusive.

This is not so for the Weinan loess section investigated recently by *Zhu et al.* [1994b]. Here the Matuyama-Brunhes and upper Jaramillo transitions have been carefully documented over stratigraphic thicknesses of about 50 cm (27 horizons in L_8) and 11 cm (13 horizons in L_{10}), respectively. The sedimentation rates are not the same over both transitions, and it is concluded that some 3200 years were required for the Jaramillo termination compared with about 5000 years for the Matuyama-Brunhes transition. These values are consistent with previous estimates obtained from marine sediments [*Clement et al.*, 1982; *Clement and Kent*, 1991] and also with the very first estimates derived from volcanic sequences more than 25 years ago [*Cox and Dalrymple*, 1967]. The actual paths defined by both sets of VGPs are strongly biased toward two antipodal longitudinal bands, one over eastern Asia and Australia, the other over the Americas. In recent years, several other studies have identified these two sectors as preferential VGP paths during reversals, and considerable debate (succinctly summarized by Zhu and his colleagues) has ensued. The Weinan record is particularly important in this context because of its geographical location, which helps to counterbalance the preponderance of data from European sites. Given that the VGP paths exhibit this important longitudinal bias, it is important to note that sequences of individual poles indicate numerous back-and-forth oscillations within, and occasionally outside, the main bands. The question of recording fidelity therefore arises. Do the oscillations represent true geomagnetic behavior, or is the possibly complex interplay of sedimentation and magnetization responsible? No definitive answer can yet be given, but *Zhu et al.* [1994b] take pains to fully investigate the mineral magnetic properties of their samples and also to consider the effects of sedimentological smoothing [*Langereis et al.*, 1992]. *Zhu et al.* [1994b, p. 157] conclude that genuine geomagnetic signals are involved, "at least in the main trends," but feel that "it is crucial that additional records be obtained." Bearing in mind that the Weinan M-B reversal results differ markedly from the corresponding Xifeng data described above [*Sun et al.*, 1993], we hasten to concur.

Milankovitch Periodicities and Magnetic Timescales

As discussed above, the susceptibility time series at Halfway House, Alaska, and Baoji, China, provide important clues in support of Milankovitch's original concept that variations in the Earth's orbital parameters are responsible for the waxing and waning of ice ages. Modern calculations [*Berger*, 1988] indicate that for the last 5 m.y. the main periodicities to be expected fall approximately at 100, 40, and 20 ka. These arise from variations in (1) the eccentricity of the Earth's orbit, (2) the obliquity of the spin axis, and (3) the precession of the equinoxes, respectively. In Alaska, *Begét and Hawkins* [1989] identified these signals by means of spectral analysis and thus provided the first conclusive evidence for the link between land-based climate proxy data and orbital forcing. This extended the earlier findings from oceanic oxygen isotope results [*Hays et al.*, 1976] which were so important in what *Berger* [1988] refers to as the "Milankovitch renaissance." The Halfway House record is rather short (~250 kyr), so it is reassuring to find that the 2.5-m.y. record at Baoji [*Wang et al.*, 1990] yields similar results, although in this case the sample spacing is such that the precession signal is scarcely detectable. An independent analysis based on grain size determinations on 1586 samples taken at 10-cm intervals throughout the Baoji section has recently been reported by *Ding et al.* [1994]. They show that variations in the size of the constituent particles of loess and paleosol exhibit systematic differences attributable to glacial-interglacial cyclic variations in the intensity of the winter monsoon winds blowing from Siberia. This extremely important data set also yields the main Milankovitch periodicities and thus provides yet another line of evidence supporting the concept of orbitally forced insolation changes controlling global climate.

The magnetic timescale derived by *Kukla et al.* [1988] is independent of astronomical tuning. As discussed above, it is based on constant eolian magnetic influx but neglects the periodic enhancement of the susceptibility signal by soil formation processes. A more reliable susceptibility timescale, however, may be developed by matching the susceptibility record (compare Figure 10) to the astronomically tuned oxygen isotope record. The transition to such a time series requires synchroneity between the climatically induced isotope changes and the largely pedogenically provoked variations in susceptibility. In comparison with a time series interpolated linearly between polarity boundaries, sedimentation rates of loess and paleosol beds are altered by this fine-tuning procedure. The Brunhes accumulation rates at Luochuan derived from the susceptibility time series tuned to the SPECMAP timescale [*Imbrie et al.*, 1984] show dramatic changes between warm and cold periods (Figure 17). On average, loess accumulates about 2.5 times faster during cold climate (11.5 ± 4.2 cm kyr^{-1}) than during times of soil formation (4.5 ± 1.0 cm kyr^{-1}). At Luochuan the highest accumulation rate exceeds 15 cm kyr^{-1}, which is markedly higher than the long-term trend throughout the Brunhes discussed above.

Increased sediment input during glacial times is also found in deep-sea cores, particularly in core V21-146 from the North Pacific [*Hovan et al.*, 1989]. It is observed that eolian flux reaches values near 600 mg cm^{-2} kyr^{-1} during cold, dry periods and falls to about a third of this during interglacials. Furthermore, this important record permits the dust flux variations over the Pacific to be correlated with the lithological changes found on the Chinese loess plateau. High dust flux over the Pacific is synchronous with loess deposition at Xifeng, whereas low values correspond to intervals of soil formation. Spectral analysis of the V21-146 dust flux record [*Hovan et al.*, 1991] yields peaks at 100, 41, and 19 ka. *Rea* [1993, 1994], in his excellent discussions concerning the geological history of wind, emphasizes the significance of these findings to global change. Correlations between oceanic and continental records are thus beginning to reveal a complete hemispheric picture, and the identification of the main Milankovitch cycles ties the whole system to the orbital forcing mechanism in a very satisfactory manner.

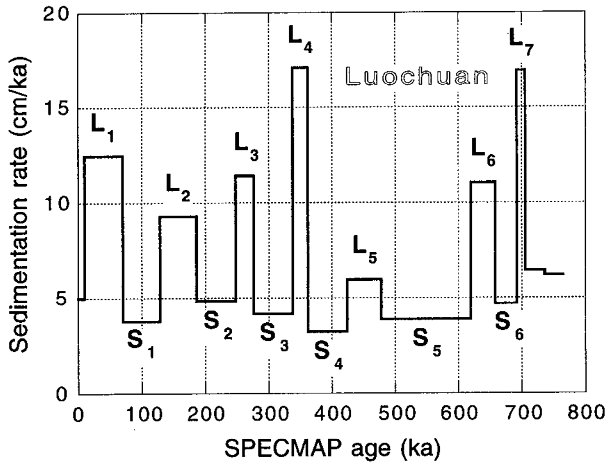

Figure 17. Loess accumulation rates at Luochuan during cold (represented by loess horizons (L)) and warm (represented by paleosol horizons (S)) climatic periods of the Brunhes epoch. The time base was obtained by matching the susceptibility series to the SPECMAP timescale of *Imbrie et al.* [1984].

Susceptibility and Paleorainfall

The ^{10}Be record in the Chinese loess at Luochuan shows features similar to those obtained from $\delta^{18}O$ records of deep-sea sediments which reflect paleoclimatic variations [*Shen et al.*, 1992]. Beryllium 10 is produced in the atmosphere and deposited on the loess plateau by precipitation and dry fallout. High ^{10}Be concentrations are associated with paleosols; low concentrations due to greater dilution are found in loess beds. These concentration variations have been matched with the SPECMAP $\delta^{18}O$ record during the Brunhes chron. Excellent agreement between the two records has been found, so that a timescale based on the ^{10}Be variations could be established [*Shen et al.*, 1992]. Using this timescale, a loess accumulation model has been developed which shows low accumulation (~5 cm kyr^{-1}) during warm climate periods and much higher accumulation rates during cold periods (10–20 cm kyr^{-1}) during the Brunhes chron. As expected, the ^{10}Be flux parallels the accumulation rates.

The ^{10}Be concentration at Luochuan during the last 140 kyr reveals a profile similar to that of the bulk susceptibility [*Beer et al.*, 1993]. The magnitude of the susceptibility variations between loess and paleosols, however, is about a factor of 2 larger than that of the ^{10}Be concentrations (Figure 18). The corresponding ^{10}Be flux can be considered to consist of a local atmospheric component F_a and a dust component F_d (Figure 19). It is mainly the dust flux component that causes high ^{10}Be flux during the stormy and cold times of loess deposition, whereas the atmospheric flux may

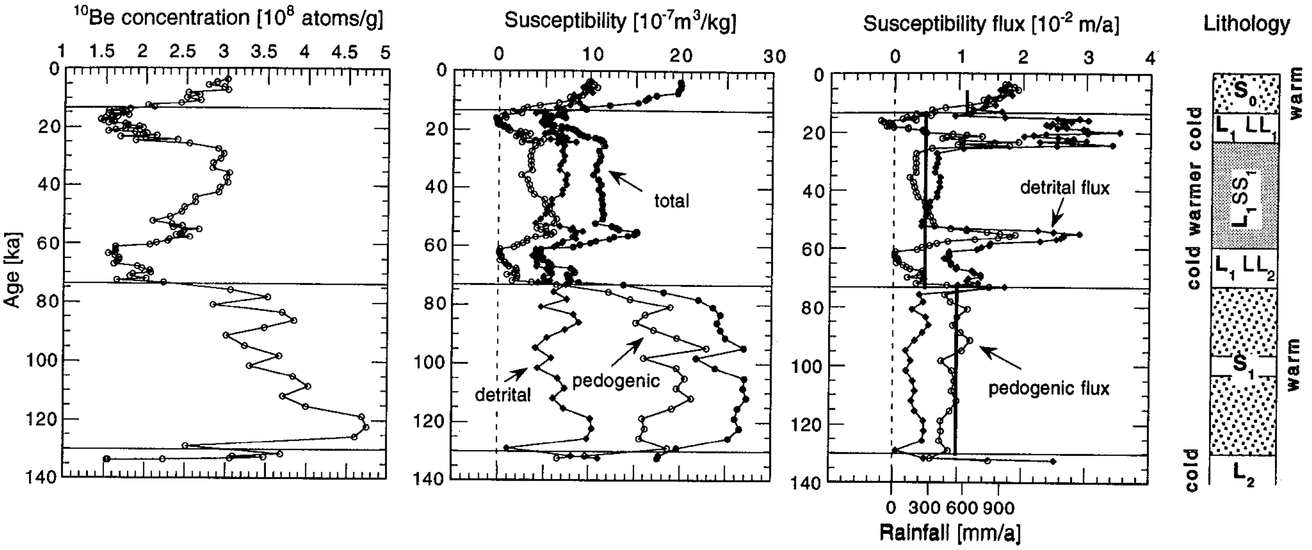

Figure 18. Beryllium 10 concentration, susceptibility, susceptibility flux and lithology at Luochuan during the last 140 kyr. The pedogenic and detrital susceptibility contributions have been calculated from comparison of the ^{10}Be and the susceptibility fluxes. Paleorainfall has been derived from the pedogenic susceptibility flux. The heavy bars of the pedogenic flux are interpreted to denote average paleoprecipitation of the major climatic periods represented by the lithological units S_0 (~600 mm yr^{-1}), L_1 (~310 mm yr^{-1}), and S_1 (~540 mm yr^{-1}).

be considered constant to a first-order approximation. The susceptibility flux (Figure 19) may be composed of a detrital (dust) component $F_{d'}$ and a pedogenic component F_x formed in situ. If it is assumed that the susceptibility dust fluxes and the ^{10}Be dust fluxes are proportional to one another, then one is able to separate quantitatively the pedogenic and the detrital susceptibility components, as shown by *Beer et al.* [1993].

Large contributions from magnetite formed in situ control the soil susceptibilities at Luochuan (Figure 18). In the Holocene soil S_0 the detrital and pedogenic components are about equal, whereas the pedogenic component of paleosol S_1 overwhelms the detrital susceptibility by far (80% versus 20%). Virtually no pedogenic components are observed in the unweathered parts (LL_1 and LL_2) of the intervening loess layer L_1. As soon as weathering becomes apparent as in the middle part (SS_1) of L_1, which corresponds to oxygen isotope stage 3, pedogenic contributions rise immediately to values of about 50%.

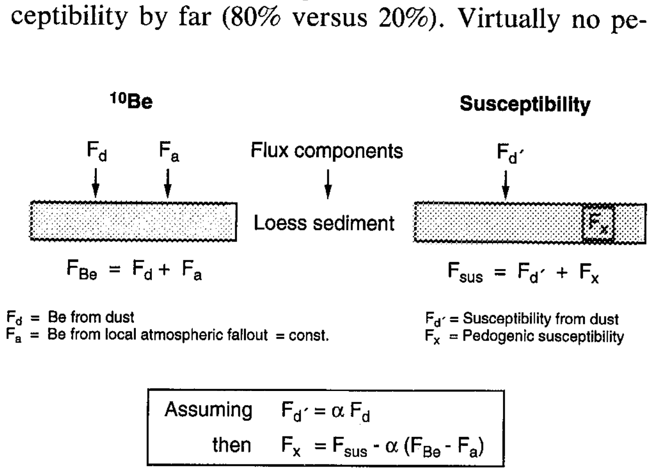

Figure 19. Model comparison of ^{10}Be deposition and susceptibility formation in loess. *Beer et al.* [1993] have demonstrated the correlation variable α between the ^{10}Be dust flux and the susceptibility dust flux to be constant to a first-order approximation.

Having quantified the pedogenic susceptibility contribution, a further observation led *Heller et al.* [1993] to estimate paleoprecipitation from the susceptibility flux magnitudes. *Liu et al.* [1992] had noted that the bulk susceptibilities of specific loess and paleosol beds were directly related to present-day mean precipitation, with the mean susceptibilities increasing linearly from low values in the drier western parts to higher values in the more humid central areas. Taking present-day precipitation as a calibration value for the Holocene pedogenic susceptibility, *Heller et al.* [1993] reconstructed the precipitation history in Luochuan. The model shows precipitation during deposition and development of paleosol S_1 similar to today, completely suppressed rainfall for the cold intervals (LL_1, LL_2) in L_1 corresponding to oxygen isotope stages 2 and 4, and short high-rainfall peaks at each of the lithological changes corresponding to major climate changes, i.e., at the oxygen stage boundaries 2-3, 3-4 and 4-5. It is not clear if these peaks represent features similar to the frequent instabilities observed in the Greenland Ice Core Project (GRIP) ice core [*GRIP Members*, 1993] or if they represent model failures which perhaps may originate from inaccurate dating.

When developing the above timescale and deriving the accumulation and precipitation rates for different climate conditions, it was tacitly assumed that sedimentation did not stop completely during warm climate periods. This certainly holds true for the Holocene, as one regularly experiences in China at the present time, but only accurate absolute age dating offers the possibility of further evidence for continuing loess deposition during earlier warm episodes. The thermoluminescence data given by *Liu et al.* [1995] clearly support ongoing sedimentation during the formation of paleosol S_1 at Weinan (38.3°N, 109.5°E), although well-defined accumulation rates still cannot be determined for this lithological unit because of experimental inaccuracy.

Recently, *Maher et al.* [1994a, b] extended the paleoprecipitation studies in space across the whole Chinese loess plateau using a slightly different approach. In disregarding ongoing accumulation during paleosol formation and assuming that the climofunction rainfall/susceptibility reaches saturation in a relatively short, but as yet unknown, interval, they simply derive rainfall rates from bulk susceptibility differences between paleosol and one of the most silty and unweathered loess beds. They calibrated their model with nine modern soil susceptibilities from sites across the loess plateau using a logarithmic relation between susceptibility and rainfall and arrive at results that are not much at variance with the analysis given by *Heller et al.* [1993].

Maher et al. [1994a] also identify dramatic changes in paleorainfall in time and space due to variations in the structure of the Asian monsoon. During interglacial periods and in the early Holocene, they see increased rainfall throughout central China, but particularly pronounced in the western plateau. During glacial periods, rainfall was reduced across the whole loess plateau, especially in the southeast.

Rainfall estimates of this kind represent an important step toward completing the overall paleoenvironmental picture. Evidence is now emerging that precipitation changes are just as important as temperature variations for assessing global change scenarios. *Stine* [1994], for example, has investigated relict tree stumps in California and Patagonia and reports that they indicate severe intervals of wetland desiccation during medieval times. He suggests that the cause was a reorientation of the midlatitude storm tracks due to a general contraction of the circumpolar vortices. Whatever the cause, however, he stresses that aberrant atmospheric circulation can produce far greater departures in precipitation than in temperature. He therefore makes the understandable plea that such well-known terms as "Medieval Warm Period" or "Little Climatic Optimum" be replaced by neutral terms such as "Medieval Climatic Anomaly." This will help to avoid prejudice in future analyses.

Early Fossil Hominids

There are many important fossil associations with loess deposits around the world, among which are the large mammoths found immediately above the Sangoman paleosol (about 100 ka) in Nebraska [*Schultz*, 1968]. But the most important example involving paleomagnetic results from loess is that of Lantian man (*Homo erectus*) in China. Two separate finds are involved, a mandible from a site ~30 km east of Xian and a cranium from a site some 20 km farther to the east. Magnetostratigraphic investigations [*An and Ho*, 1989; *Shaw et al.*, 1991] indicate clearly that the mandible falls in a normal polarity zone, whereas the cranium occurs lower down in a reversed zone. Correlation of the magnetic results and the loess/paleosol alternations with the Luochuan section indicates ages of ~0.65 Ma and ~1.15 Ma for the mandible and cranium, respectively. The Lantian cranium thus appears to be the oldest reliably dated Chinese hominid site.

CONCLUSIONS

Two broad lines of enquiry have emerged from magnetic investigations of loess: magnetostratigraphy and paleoclimate. The first relies on the recording of geomagnetic polarity by some form of remanence acquisition; the second is due to the connection between pedogenesis and susceptibility. In the case of remanence, great thicknesses of loess, particularly in China, yield superb sections with the Brunhes, Matuyama, and uppermost Gauss chrons clearly identified. In this way, the time represented by loess deposits has been drastically revised, from ~1.2 m.y. to ~2.5 m.y. Similar conclusions hold for Alaska and Argentina, even though the sections involved are 10 times thinner. One of the sections studied in China contains an important early hominid cranium, which can now be dated paleomagnetically to ~1.15 Ma. Despite these and other significant findings which magnetostratigraphy provides, the actual mechanism by which the sediments acquire their remanence remains uncertain. Some sites have apparently retained a detrital signal from their time of formation and consequently provide high-resolution records including details of the order of centuries, within the Blake subchron for example. Other sites carry a chemical remanence apparently acquired over intervals of the order of 10,000 years, leading to considerable smoothing and loss of detail. There is a pressing need for more systematic study.

Understanding, and exploitation, of susceptibility data is more advanced than is the case for remanence. The correlation between susceptibility and climate is firmly established. In China and central Asia, high susceptibilities correspond to warmer and wetter conditions, which are convincingly correlated with the deep-sea oxygen isotope record. In Alaska and Argentina, however, the two signals are anticorrelated. Two separate models are responsible. Where correlation is positive, susceptibility is enhanced by pedogenic processes, which probably include production of ultrafine (<100 nm) Fe_3O_4 particles by magnetotactic bacteria. In the case of negative correlation, local meteorological factors such as strong and turbulent winds able to entrain large (~50 μm) Fe_3O_4 grains are more important.

At certain sites in Alaska and China, power spectra obtained from susceptibility time series exhibit strong peaks at ~100, ~40, and ~20 ka corresponding to the insolation periodicities predicted from astronomical theory, further fueling the so-called Milankovitch renaissance. Moreover, it has proved possible to combine the susceptibility data with measurements of ^{10}Be to estimate the paleorainfall in north central China since the last interglacial. Thus the various loess profiles are starting to provide valuable paleoenvironmental information germane to the global change debate. Indeed, the very timing of loess deposition, now shown to have commenced close to the Gauss-Matuyama boundary (2.6 Ma), coincides with a widely documented global climatic shift and accelerated uplift of the Tibetan plateau.

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