Loess magnetism, environment and climate change on the Chinese Loess Plateau

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Loess Magnetism, Environment and Climate Change on the
Chinese Loess Plateau

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Abstract

The strain on the Earth’s natural resources has reached dimensions, which are denounced by many people as reckless and indifferent towards our environment and our future. Therefore the impact of newly developed technologies on the environment and on our society needs to be thoroughly investigated and to be scrutinised ethically. In particular, the global climate is endangered by the steady increase of energy consumption. The influence of human activity on climate can be evaluated by comparing with the climate history of our planet before the Holocene. Different climate archives such as ice cores, marine and lacustrine sediments or loess/palaeosol deposits may be studied and compared world-wide in order to recognise global or regional climate changes in the geological past.

The Chinese Loess Plateau (CLP) contains one of the largest continental climate archives. Preferentially during glacial times, huge loess volumes with thicknesses up to 300 m have been accumulated there since 2.6 Ma. These deposits do not consist of a uniform pile of loess only; they are interlayered by fossil soils (palaeosols), which were formed during interglacial periods. The present study focuses on a magnetostratigraphic dating problem that occurs if loess/palaeosol sediments are compared with other global climate archives, such as marine sediments. Each palaeosol horizon of the Chinese loess/palaeosol sequences correlates with an odd-numbered marine oxygen isotope stage (MIS). The Matuyama/Brunhes polarity boundary (MBB) occurs in marine sediments within MIS 19, representing an interglacial time interval, which corresponds to palaeosol S7 in Chinese loess/palaeosol sediments. The MBB, however, is observed in the underlying loess horizon L8, representing a glacial period.

In order to shed some light on this problem, one of the thickest loess/palaeosol outcrops of the central CLP – the loess/red clay section of Lingtai (north-west of Xi’an) – has been investigated palaeomagnetically. The outcrop was sampled continuously resulting in the enormous amount of ~12’000 oriented samples of 8 cm³ volume. They were carefully extracted and prepared by Prof. Yue Leping and his co-workers from the Northwest University of Xi’an. The low field susceptibility and its frequency dependence of all samples has been measured in order to characterise the general lithostratigraphy and the palaeoclimatic significance of the loess/palaeosol sequence. These magnetic proxy data reflect essentially the lithostratigraphy, which was already recognised in the field.
Palaeosols are well distinguished from loess horizons by enhanced susceptibility values due to enriched ferromagnetic minerals. The enhanced frequency dependence of the palaeosols indicates the presence of superparamagnetic minerals, too.

The characteristic natural remanence (ChRM) direction throughout the MBB transition interval has been investigated magnetostratigraphically in detail at Lingtai and in the nearby section at Baoji (80 km). The MBB has been observed in both outcrops within the loess L8. The ChRM directions jump several times between reversed and normal polarity in both sections, but pattern and number of jumps differ. Further inconsistent magnetostratigraphic MBB transitions have been observed in the Weinan loess section (central CLP) and in the ODP core 792A near Japan. The resulting virtual geomagnetic pole paths are not well defined. They change polarity several times and do not prefer specific longitudinal bands. Hence, the MBB transitional behaviour in loess may not reflect geomagnetic field variations, but may rather results from a complex natural remanence acquisition process. In addition, the MBB record is delayed and downwards shifted in the Chinese loess/palaeosol column by 1.5 to 2 m corresponding to about 25'000 years.

Rock magnetic properties of loesses and palaeosols have been investigated in order to determine and quantify the mineral phases carrying the signal of the natural remanent magnetisation (NRM) in the MBB transition interval. Generally, four magnetisation components contribute to the acquisition spectra of isothermal remanent magnetisation (IRM) of different grain size fractionated loess/palaeosol samples. Two components, $D_1$ at 80 mT and $D_2$ at 110 mT, dominate the bulk remanence of the loesses, whereas a low coercivity component $P$ at 30 mT contributes mainly in palaeosols. Both, loesses and palaeosols have a high coercivity component $H$ at 550 to 760 mT. All four components have been quantified using a linear mixing model. Component $P$ is due to pedogenic maghemite as indicated by a Curie temperature of 620 °C and low coercivity and because its contribution to the bulk remanence increases with pedogenesis. High coercivity, low magnetisation contribution and a Curie temperature of 680 °C of component $H$ indicate a hematite remanence carrier. $H$ is of detrital origin, because its contribution to the bulk remanence is constant in all analysed loess/palaeosol samples. The thermomagnetic curves of grain size fractions carrying mainly components $D_1$ and $D_2$ show Curie temperatures of 610 °C and identify maghemite as remanence carrier. This maghemite differs
from that of component P by grain size and by oxidation degree (higher coercivity). The maxima of the coercivity spectra and the thermomagnetic behaviour of D1 and D2 show small but significant differences. Component D1 is interpreted to be of detrital origin because it contributes fairly constantly to the bulk remanence of different loesses and palaeosols and dominates the bulk remanence of pristine loesses with contributions of just more than 50%. Component D2 dominates in weathered loesses such as loess L8, with contributions of about 60% and contributes variable remanence amounts in different loess/palaeosol samples. The dispersion of the magnetisation contribution of D2 in the different samples is twice than that of D1.

The lock-in model of Bleil and von Dobeneck (1999) that explains the delayed NRM acquisition has been applied and developed further to solve the magnetostratigraphic conflict of the MBB at Lingtai. The model comprises the coexisting variable remanence components present in loesses and palaeosols. It is assumed that first a detrital remanence component is acquired and later a pedogenic remanence component. The contribution of both components has been established experimentally by stability tests of the anhysteretic remanent magnetisation in the samples continuously taken across the MBB transition. The lock-in model does not only explain the MBB shift of 1.74 m (equivalent to 22’000 years), it also explains successfully the observed polarity jumps in zones where detrital and pedogenic magnetisations interfere in a complex manner.

The observed and the modelled Matuyama/Brunhes transitional magnetisation records give evidence that the natural remanent magnetisation in loess is much later acquired than that in deep sea sediments. Hence, the MBB has been displaced downwards into loess L8 whereas the real stratigraphic position has to be looked for in the overlying S7. The apparent magnetostratigraphic contradiction of the correlation of the loess susceptibility data with the marine oxygen isotope data has been solved. In addition, the natural remanent magnetisation of loess sediments does not reflect geomagnetic field behaviour during the MBB polarity transition. Their directions in fact are dominated by variable contributions of detrital and pedogenic remanence carriers.
Zusammenfassung


Um diesem Problem nachzugehen, ist eine der mächtigsten und komplettesten Löss/Paläosolabfolgen des zentralen chinesischen Lössplateaus – der Löss/Red clay Aufschluss von Lingtai, nordwestlich von Xi’an – untersucht worden. Der in etwa 300 m mächtige Aufschluss wurde kontinuierlich beprobt. Rund 12000 Einzelproben wurden orientiert entnommen und durch Prof. Yue Leping und


Mit detaillierten gesteinsmagnetischen Untersuchungen wurden die Remanenzträger von Lössen und Paläosolen analysiert. Anhand von Aufmagnetisierungsspektren der isothermalen remanenten Magnetisierung (IRM) von Korngrössenfraktionen verschiedener Lössen und Paläosole konnte gezeigt werden, dass im wesentlichen vier Komponenten zur remanenten

Um die Diskrepanz in der stratigraphischen Position der Matuyama/Brunhes-Grenze zu erklären, wurde ein Modell von Bleil und von Dobeneck (1999), dass den verzögerten Erwerb der natürlich remanenten Magnetisierung erklärt, auf Lingtai angewendet und weiterentwickelt. Bezugnehmend auf die koexistierenden variablen Remanenzkomponenten in Lösse und Paläosolen zeigt das
Modell auf, dass zuerst eine detritische Remanenzkomponente erworben wurde und danach eine pedogene Komponente. Der Anteil dieser Komponenten in den kontinuierlich entnommenen Proben des MBG-Übergangsbereiches wurde experimentell durch Stabilitätstests der anhysteretischen remanenten Magnetisierung gekennzeichnet. Das Modell erklärt nicht nur den Versatz der MBG um 1.74 m (entspricht 22 000 Jahren), sondern modellierte erfolgreich die beobachteten Polaritätssprüinge in den Bereichen, in denen detritische und pedogene Magnetisierung in komplexer Weise interferieren.

Part I

Introduction
1. Research objectives

In August 2002 the news about catastrophic floods in the world did not seem to end: Water devastation in Bangladesh, landslides after heavy rain in Iran and China, floodings in Germany, Austria, the Czech Republic, and Nepal, thunderstorms on the Philippines, cloudbursts over the Himalaya and extreme aridity in Italy, Vietnam and the Canadian Midwest, a Croatian river that flows in opposite direction due to flooding. These were the news of two days only this year (NZZ, 13./14. 08. 2002). And the longer the more, the obvious question must be asked how much all these “natural” disasters are related to anthropogenic influence on the Earth’s atmosphere, biosphere and hydrosphere.

Just two examples of human influence: Condensation trails of aircrafts directly influence the diurnal temperature variation. The mean difference between maximal day and minimal night temperature in the United States (as measured at 4000 weather stations) was 1.1 °C higher between September 11th and 14th 2001 than the mean variation of the analogue three day period between 1971 and 2000, (Travis et al., 2002). The present climate over southern Asia is another example. A 3 km high brown cloud consisting of ash, grime, acids and other harmful substances reduces the insolation by about 10 to 15 %. Lower surface

![Figure 1.1: Temperature anomaly of the northern hemisphere mean annual surface temperature from the calibration period between 1901 and 1980 (after Mann et al., 1999). Systematic instrumental measurements started in 1860.](image-url)
temperatures and concomitant warming of the atmospheric layers above the dirty cloud changed the regional climate and the monsoon rhythm (UNEP and C4, 2002).

Indeed, human influence on the global climate can not be neglected anymore. But we have to keep in mind that the Earth’s climate can change also without human influence. Instrumental temperature measurements (Figure 1.1) indicate that the mean surface temperature increased by about 0.5°C from 1860 to the present. This record is not long enough to discriminate if this warming is due to natural climate variations or human activity. In addition, it has to be taken into account that the global climate does not respond linearly to impacts (anthropogenic, natural disasters) and changes can happen suddenly. It is an intrinsic property of the climate to be chaotic and erratic forecast solutions are found in deterministic physical models if the input parameters are not precisely known. If we want to predict future climate scenarios, both, the natural and the anthropogenic aspects have to be considered. Palaeoclimatic data can be used to extend the climate record and to provide a longer time frame (hundreds to tens of thousands of years) for evaluating the significance of the warming of the last 140 years.

The Earth’s past climate was subjected to much larger variations than shown in Figure 1.1. Billups and Schrag (2002) measured magnesium/calcium ratios to show that the seawater temperature was on the average 2 °C higher during the Miocene than during the Pleistocene/Pliocene. Since the early Eocene the Earth’s climate cools down. The beginning of this cooling trend is marked by first ice sheets in Antarctica 36 Ma ago (Miller et al., 1987). This long-term cooling brought the Earth into a critical state, which enabled the growing of ice sheets on the Northern Hemisphere as well. Driscoll and Haug (1998) assumed that the closure of the Isthmus of Panama caused a new thermohaline circulation in the North Atlantic 4.6 Ma ago. More moisture has been brought to the Eurasian continent and further transported eastwards to Siberia via westward winds (westerlies). This moisture discharges through the large Siberian drainage system (~ 10 % of the global river discharge) flowing toward the Arctic (e.g. rivers Yenissey, Lena, Ob). As a consequence, the Arctic Ocean has been enriched with freshwater which facilitated the formation of sea ice. In addition, the Pliocene uplift of the north-western Tibetan Plateau caused a reorganisation of atmospheric circulation over Siberia 4.5 Ma ago, which probably triggered the Northern Hemisphere glaciation, too (Ding et al., 1992).
Plio-Pleistocene to Holocene Loess/palaeosol sediments are amongst the best continental recorders of palaeoclimatic changes. They offer a unique possibility to study natural variations of the past global climate as in glacial times loess has been deposited and during the intervening interglacials palaeosols have formed in the aeolian sequences.

Amongst other goals, this work is intended to contribute to the complex palaeoclimate questions. It tackles a major conflict in the magnetostratigraphy of Chinese loess/palaeosol sections: the position of the Matuyama-Brunhes geomagnetic polarity boundary (MBB). If loess/palaeosol sequences are claimed to be archives of global climate, then they have to be comparable to other climate archives such as oxygen isotope records of ice cores or marine sediments. An accurate timescale has to be established, but exactly here major controversies arise when considering the position of the MBB in the famous loess/palaeosol sequences of the CLP. The boundary has been observed in the marine oxygen isotope stage (MIS) 19 a horizon, which corresponds to an interglacial period. In most of the Chinese loess/palaeosol sequences, however, this major polarity change occurs in a loess layer, which clearly represents a cold climate interval. This causes a large age discrepancy for the MBB of about 23000 to 26000 years between magnetostratigraphic records on land and in the marine realm.

In order to provide some answers to this problem, one of the most complete loess/palaeosol sections of the central CLP – the loess-red clay sequence of Lingtai – was sampled continuously and studied in detail for a thorough palaeomagnetic analysis. We concentrated on the following points:

- Low field susceptibility measurements for a general lithological–palaeoclimatic characterisation.
- Magnetostratigraphic characterisation of the whole sequence to provide age tie points.
- Detailed magnetostratigraphic and rock-magnetic investigations of the Matuyama-Brunhes boundary.
- Detailed analysis of rock-magnetic and chemical properties of typical loess layers and palaeosols for palaeoclimatic interpretation.
- Combination of results and solution of the contradictory evidence of the position of the MBB position in marine and loess/palaeosol sediments by modelling the remanence acquisition process in Chinese loess/palaeosol sediments.
2. Introduction
2.1 Philosophical introduction

As time plays an important role in this work, different aspects of time are discussed in the following paragraphs.

2.1.1 Philosophical time

How often do we say: “I had no time”. This statement implies the possibility of “having time”. But what do we mean? Can we really have time and what is time? In order to answer this question we have to define the term *time*.

Time may be defined as a concept allowing evolutionary processes or sequences of events to happen. These events do not happen over time, they themselves represent the time. Before classifying *time*, two other questions arise: Can we (somehow) feel time? Can we use time? The answer to the first question is *yes*, but we can not feel time directly, as human beings do not have a time sensing organ. We can feel rhythms as do plants or animals. Secondly, we cannot use the time in a general sense but we can manipulate time under certain conditions. Thus, time can not be termed generally and each evolutionary process has its own time. Such a concept is best described by the word “phenomenon”, which has two meanings: 1. Something presenting itself to our senses. 2. Something extraordinary or unusual. Both meanings are appropriate.

We have an intrinsic impulse to try to understand natural phenomena. As knowledge much was smaller at the beginning of human incarnation, inexplicable phenomena were attributed to actions of different deities. With the transition from the primordial society to the slave holder society, natural phenomena were studied in more detail and explained in a scientific and philosophical way. Egyptian scholars were busy in creating calendars already around 5000 B.C. The importance of time was so eminent during the antiquity that specific deities personalising time were introduced. Chronos for instance, was the Greek godhead of time and Thot the Egyptian equivalent, responsible for scheduling the Egyptian calendar. Early philosophers kept track on developing more sophisticated scientific descriptions of the phenomenon time. The trivial words *past, present* and *future* were introduced first by Parmenides of Elea (514-440 B.C.). The philosophy of the Elean school – led by Thales of
Milet (624-546 B.C.) – was coupling time with motion, which caused sophism. Aristoteles (384-322 B.C.) for instance, held up the thesis that every moving body has a natural tendency to come to stagnancy. Much later, this was disproved by Isaac Newton (1643-1727) who introduced friction into motional processes. An important contribution to the separation of time and motion was already achieved by Galileo Galilei (1564-1642) when he proposed that bodies fall with the same velocity independently of their mass, if air resistance is neglected. Newton discarded the time conception of the antiquity and introduced an absolute time, which flows by itself. This opinion was disproved essentially by Albert Einstein’s (1879-1955) relativity theory. Einstein actually answered the time question in a facetious way: “Time is, what we can read off the clock.”

Two fundamental elements control the occurrence of evolutionary processes: space and time. If only one of these phenomena would occur, evolution will not start: If there is no space, where do sequential time events occur? And, if there is space without events occurring (which represent the time), nothing will happen. Thus we can consider space and time as continuum. This model may not apply to stationary processes when past, present and future are indistinguishable. Such steady-states may occur below the Planck-dimensions (time < 5.35x10^-44 s and length < 1.616x10^-35 m) in singularities such as moving at light speed or before the big bang. Nevertheless, most of the macroscopic (above the Planck dimensions) phenomena can be described by that time model.

2.1.2 Physical time

The wish and the necessity to measure time is as old as human existence and is linked to the basic activities of our life like hunting, seeding or harvesting. Religious traditions established chronologies in order to reinforce religious events. Time measurements started in the antiquity observing natural events like lunar phase re-occurrences, the rotation of the Earth around its axis and around the sun. At about 3000 B.C. the Egyptians invented oil dials in which oil was burning down to specific markers of the oil reservoir. Gnomony, the science of construction and calculation of sun-dials, was also developed quite early. The term Gnomon is translated from the Greek γνώμον (shadow bar). The Egyptians building earliest sun dials were familiar with the relation between the length of the shadow and the time of the day at about 1500 B.C. (Zenkert, 1984). Water
dials were used during Socrates (469-399 B.C.) times to limit the speaking time at public gatherings. During the early medieval ages the interest of more accurate time determination emanated from monasteries. The monks lived along strong disciplinary rules and needed a timed life rhythm. Water and sun dials have been developed further to perfection at that time.

The invention of the mechanical clock in Europe is not well dated. The first clocks with weights may have been built around 960 A.D. These clocks had still the escapement problem, which was tackled about 300 years later by the first professional clockmakers. The division of one hour into 60 minutes and one minute into 60 seconds is well documented in 1345. With the scientific revolution – heralded by Nicolaus Copernicus (1473-1543), continued by Johannes Kepler (1571-1630), Galileo Galilei (1564-1642) and Isaac Newton (1643-1727) to name only some of them – the need of more precise clocks grew. Inventors like Cristian Huygens (1629-1695), John Harrison (1693-1776) and Thomas Mudge (1715-1794) hold for the further development and accuracy of the mechanical clock. The time of the electrical clocks began when Alessandro Volta (1745-1827) invented the first dry battery in 1801. During the 19th century time could already be measured so accurately that Léon Foucault (1819-1868) determined the speed of light. He obtained a value of 298 000 000 m/s in 1862. This is only ~ 0.6 % smaller than the present day value of 299 792 458 m/s. Modern time measurements rely on atomic clocks. A second is nowadays defined as the duration of 9 192 631 770 periods of the radiation corresponding to the transition between the two hyperfine levels of the ground state of the caesium 133 atom (13th Conference of Weights and Measures, 1967). Hence, the present day atomic clocks have an imprecision of one second in 2.5 Myr.

The construction of more and more precise clocks in the beginning of the renaissance has certainly influenced philosophers and scientists like René Descartes (1596-1650) and Isaac Newton who developed a rationalistic, entirely mechanistic world-view. When Newton postulated absolute time and space, Gottfried Wilhelm Leibnitz (1646-1716) challenged these postulates and tried to combine the rational world-view of Descartes with Christian belief. His studies of the infinitesimal geometry led to the development of the new calculus in 1675 and opened the route to the solution of differential equations. These equations can – to a certain extent – predict future behaviour of complex physical processes like weather or climate.
A physical process is defined clearly, if it is repeatable at any time and at any
time in space, under the same boundary conditions. In this way, it is considered
a physical phenomenon and ought to become an object of scientific
investigation. Newton’s apple should fall down toward the Earth’s surface today
and tomorrow at any place on Earth.

Simple ordinary differential equations can be used to predict evolutionary
processes. The following model is the numerical analogue of a non-linear
differential equation that describes the limited growth of a population (e.g.
bacteria) also called logistic growth (May, 1974, 1975):

\[ y_{n+1} = \rho y_n \left(1 - \frac{y_n}{k}\right) \]  \hspace{1cm} (2.1)

Figure 2.1: Centre: Stability diagram of the solutions of \( u_{n+1} = \rho u_n (1 - u_n) \) for different growth
rates \( \rho \). (a) The first solution \( u = 0 \) is stable for \( 0 < \rho < 1 \). (b) Above \( \rho = 1 \), stability changes to
the second solution \( u = (\rho - 1)/\rho \), which is stable for \( \rho < 3 \). The second solution becomes
unstable above \( \rho = 3 \), but shows a certain degree of order. Two and four periods occur in (c)
and (d), respectively. No regular solution is found for \( \rho > 3.57 \) – chaos occurs (see text).
The parameter $\rho$ is called intrinsic growth rate and $k$ a saturation parameter that limits the growth. Scaling of $y_n$ with $u_n = y_n/k$ simplifies the above equation to:

$$u_{n+1} = \rho u_n (1 - u_n)$$  \hspace{1cm} (2.2)

The two equilibrium solutions are:

$$u_n = 0, \quad u_n = \frac{\rho - 1}{\rho}$$  \hspace{1cm} (2.3)

**Figure 2.2:** Solutions of $u_{n+1} = \rho u_n (1 - u_n)$ for equal $\rho$ ($\rho = 3.7$) but slightly different start values ($u_0 = 0.3$ and $u_0 = 0.301$). Small uncertainties of the initial conditions have big implications for the long-term behaviour of the population evolution. A long-term prediction is impossible. This effect is called *Butterfly effect* and was first demonstrated by the meteorologist E. Lorenz (1963). N.B.: The prediction is repeatable if the same initial conditions are chosen.

Figure 2.1 shows the long-term stability of both solutions for different growth rates $\rho$. A small growth rate leads to a decrease of the population members. Regarding long times, the population dies. If $\rho$ is between 1 and 3, the population grows and the number of members is constant. Between $3 < \rho < 3.57$ the solution is unstable and the number of members varies periodically. Too high growth rates result in chaotic long-term behaviour. The number of the population members varies without any regularity. At this point it is very difficult to make long-term predictions of the evolution behaviour if the initial conditions are not known precisely enough (Figure 2.2).
Hence, evolutionary processes can be described by deterministic scientific regularities, but their long-term prediction is very sensitive to the accuracy of the initial conditions. The prediction of complex systems may be best described by the title of the talk, which Edward Lorenz gave at a meeting of the American Association for the Advancement of Science in Washington, D.C. in 1972: “Predictability: Does the Flap of a Butterfly's Wings in Brazil set off a Tornado in Texas?”

2.1.3 Geological time

If we look around, we can discover many things that change permanently through time and space. The weather changes. The seasons change. The lunar phases change. Many species of the Earth’s flora and fauna have flourished and have withered. And we? Our soul for instance has highs and lows. Civilisations have grown and have died. After all, human beings are only sojourners on this planet. Solely oceans and rocks seem to remain unchanged since we started thinking. But is this really true? Xenophanes (570-480 B.C.) asked the same question. When he was sitting on a mountain looking around, he found eventually shells at his feet, shells, which he had seen only at the beach before. How did they come to the mountain? Where from could they come? Did children forget them when playing? But the shells were also within the rocks nearby. Had the sea once covered the mountain or ascended the mountain? Nowadays, of course, we know that rocks and oceans are subject to changes, too. But these changes require much more time than we can imagine – geological times.

Many geological processes lead to changes of the Earth’s crust. Therefore rocks are important witnesses of the Earth’s geologic past. Geological stratigraphy deals with sedimentary rocks, which were formed at the Earth’s surface, usually in horizontal layers. Based on the assumption of horizontal layering of sediments Nicolaus Steno (1638-1686) formulated the principles of stratigraphy in 1669. The most important one is that bottom layers are older than top layers in undisturbed sediments. Using Steno’s principles, geologists were able to construct a relative time scale. Different outcrops could be correlated by using index fossils present in layers of different sections or by comparing similar
lithologic layers. The relative chronology nowadays can be put on an absolute time scale using radiometric dating.

About 4.6 Gyr of Earth’s history are difficult to imagine when compared to human history. If the age of the Earth is converted into the equivalent of one day, the genealogical tree between human beings and monkeys would split at 23:58:07 and Homo sapiens would develop since 23:59:45 – a single life time would be only ~ 1.5 ms.

With the awareness of William Gilbert (1544-1603) that the earth has a magnetic field, its mathematical description by Carl Friedrich Gauss (1777-1855) and the discovery of remanent magnetisation in rocks by Alexander von Humboldt (1769-1859), magnetism also becomes interesting for geological dating. Bernard Brunhes (1867-1910) was the first to observe that the geomagnetic field reverses polarity. In 1906 he found ancient lava flows imprinted with a magnetic polarity opposite to that of younger and older rocks and concluded that the magnetic field of the Earth must have been reversed at the time of extrusion. Motonori Matuyama (1884-1958) confirmed Brunhes’ results. In the paper On the Direction of Magnetization of Basalt (1929) Matuyama showed that the polarity of the remanent magnetisation in some volcanic rocks of Plio-/ Pleistocene age depends on the age of the rock and concluded that the Earth's magnetic field must have undergone periodic reversals. The discovery of polarity changes of the Earth’s magnetic field and their fossilisation in rocks led – in combination with radiometric and palaeontological dating – to a new absolute age dating tool for the last 150 Myr: magnetostratigraphy was born.
2.2 Thematic introduction
2.2.1 Loess

The term *loess* is probably derived from the Alemannic word “losch” indicating a loose and unconsolidated material. It appeared in the scientific literature during the 19th century (Russel, 1844; Lyell, 1847; Virlet d’Aoust, 1857; von Richthofen, 1882) and characterises an aeolian terrestrial silt deposit, generally of Quaternary age. About 10 % of the Earth’s surface is covered with loess and loess-like (reworked loess) deposits. Loess can be found in semiarid and temperate climate zones (Figure 2.3). The main loess deposits are located on plains (Pampean Plain, Russian Plain), on plateaus (Chinese Loess Plateau) and along river basins (middle Rhine Basin, Danubian Basin, Mississippi Basin, middle Yellow River Basin).

![Worldwide loess distribution](image)

**Figure 2.3:** Worldwide loess distribution (after Catt, 1986; Pye, 1987).

50 to 70 % of the loess volume is dominated by silt grains of 10 to 50 μm diameter. This grain size fraction is composed mainly of quartz (40 to 60 %), feldspars (5 to 20 %), calcite and dolomite (2 to 25 %), micas and chlorite (4 to 10 %) and heavy minerals (1 to 6 %). The coarser fractions consist mainly of quartz. The clay fraction contains mainly illite, kaolinite, calcite, chlorite,
vermiculite, montmorillonite, smectite and goethite. In addition there are small amounts of other iron oxides or hydroxides and aluminium hydroxides, pyrite and mafic silicates (Liu et al., 1985). The presence of carbonates is an important factor for its texture. There are two types. Primary carbonates (calcite and dolomite) were transported together with the other minerals. Secondary carbonates are due to hydrogen carbonatic ground water, precipitation, microorganisms and weathering of calcareous shells and anorthite. Secondary carbonates often form concretions called “Lößkindel.”

The formation of loess takes place in two ways, mainly. Silty material can be produced by glacial abrasion, and is then transported by wind over large distances. Silt can be also a weathering product of salt and frost weathering in cold and arid deserts. Both processes take place mainly during glacial times.

The colour of loess is pale yellow and caused by limonite, a mixture of hydrated iron oxides such as goethite (α-FeOOH) or lepidocrocite (γ-FeOOH). After transportation and final deposition the loess is subjected to structural changes (hydroconsolidation). The overburden produces an increasing pressure which can cause collapse of the mechanically unstable loess. The collapse may occur at depths between 0.4 and 0.8 m in an artificial silt-clay mixture containing 10 % kaolinite (Assallay et al., 1998). Beside the clay content, another important collapsibility factor of loess is the pore size. Suzuki and Matsukura (1992) measured pore size distributions of different loesses and found two major pore sizes, ~ 0.05 μm and 10 μm in sandy loess. In clayey loess, pore sizes of 0.05 to 0.08 μm, 5 μm and 10 μm dominate. Above all, these pores are responsible for the collapse of loess (Osipov and Sokolov, 1995).

2.2.2 Palaeosols

Loess is a glacial sediment. During warmer (interglacial) climate periods the loess is exposed to new environmental conditions such as increased temperature, higher precipitation and biologic activity. Thus the loess may be progressively transformed into a soil (pedogenesis), that may be later covered by a succeeding loess layer. These embedded soils are called palaeosols (Ruhe, 1956). On the Chinese Loess Plateau (CLP) the changing past climate produced alternating layers of loesses and palaeosols, corresponding to cold, arid and warm, humid intervals, respectively.
During the process of pedogenesis, physical and chemical weathering cause characteristic changes in the particle size distribution and mineral composition. Easily soluble salts are removed quickly by leaching. As long as calcite is present in the soil, the pH-value is slightly alkaline. Acidification and - as a consequence - hydrolytic weathering of silicates starts as soon as the soil has been decalcified. In addition, the bi-valent iron ions of ferruginous minerals are oxidised causing the brown or reddish colour of the soils. New minerals or secondary weathering products develop from primary minerals. Only the most resistant minerals of the bedrock remain.

Soil types that develop mainly on loess substrate are for instance luvisol in moderate climate or chernozems in rather continental and arid climates (steppe). A luvisol is characterised by a small humic horizon (A) on top followed by a shallow horizon (E) containing less clay than the underlying thicker brown Bt-horizon. Often crusts of secondary calcite are formed at the transition from the Bt-horizon to the bedrock (C-horizon). Translocation of clays from the E- to the Bt-horizon is typical for this soil. Black soils are wide spread on the Russian plain. They are called “Chernozems” from the Russian чёрная земля (= tschjornaja zemlja). They have a characteristic thick (~ 1m) A-horizon over the C-horizon. This soil is common in a steppe environment with dry warm summers and cold winters. The strong continental annual climate temperature changes cause intensive bioturbation. The upper horizon is intensively stirred and extended, because the soil animals move downwards to avoid aridity and coldness during summer and winter, respectively. Hence, organic material is transported downwards, whilst limy material of the C-horizon is transported upwards (Sticher, 1997).

Shortly after burial, the palaeosols may be considered as nearly closed chemical systems (Driese et al., 2000). However, burial diagenesis can change the characteristics of a soil by oxidation of organic carbon, illitisation of smectites, dehydration and re-crystallisation of iron and manganese oxyhydroxides (goethite and ferrihydrite may dehydrate to hematite (Catt, 1995)) and pedogenic (secondary) calcite may be dolomitised.
2.2.3 The Chinese Loess Plateau (CLP)

Loess in China is distributed mainly in three different regions covering an area of 380,840 km² (Liu et al., 1985). In the north-western part of China, loess is spread between the Altay and Kunlun mountains with thicknesses of < 50 m. Similar thicknesses are reached in the north-eastern part of China, where loess

![Map of the Chinese Loess Plateau](image)

**Figure 2.4:** Map of the Chinese Loess Plateau, showing loess distribution, source regions and main wind directions. Minimal daily temperatures for December, January, February and maximal daily temperatures for June, July, August averaged for the period from 1951 and 1988 are indicated for several localities. The precipitation values are for the same periods (redrawn from Wang and Song, 1983; temperatures and precipitation data from Tao et al. (1997) except for Lingtai (Ding et al., 1999)). The main wind direction is caused by a stable anticyclone centered over southern Siberia and Mongolia from December to March (see Fig. 2.6).
occurs between the ranges of high and low Hinggan. The greatest loess thicknesses are attained in north-central China along the middle reaches of the Huang He (Yellow River). This area covers 275 600 km² (72.4 % of the total loess area in China) and is called the Chinese Loess Plateau (CLP). It can be divided into a western and a central sub-region (Figure 2.4). The western CLP is delimited in the north by the Tengger and Badain Jaran sand deserts, in the west by the Quilian mountains and in the south by the Qinling mountains. The western CLP loess formation has been influenced strongly by cold winter monsoon winds blowing through the Hexi Corridor (Lei and Sun, 1984) resulting in thick loess deposits (> 300 m). The central part of the CLP extends from the Liupan mountains in the west to the Luliang mountains in the east and from the Qinling mountains in the south to the Mu Us desert in the north. This sub-region was rather affected by the moister and warmer East-Asia summer monsoons than by the cold winter monsoons (Ding et al., 1999) and the loess thickness is smaller (mainly between 100 and 200 m). The present day climate confirms the supposed climate differences in both regions (see Figure 2.4). The source areas for the CLP are the Tengger Desert, Badain Jaran Desert, Mu Us Desert, Ulan Buh Desert, Hobq Desert (Sun, 2002) where the annual temperatures vary largely. The north-western main wind direction is controlled by the west-Siberian anticyclone.

2.2.4 The age of Chinese loess

Beside lithostratigraphy, magnetostratigraphy is amongst other methods such as \(^{14}\)C, fission track and thermoluminescence dating the most reliable dating tool for loess/palaeosol sections. The ages of loess sedimentation in China can be estimated by correlating of the observed polarity patterns of the loess/palaeosol sections with the geomagnetic polarity timescale (Cande and Kent, 1992; McDougall et al., 1992; Spell and McDougall, 1992). The Brunhes normal polarity, Matuyama reversed polarity and Gauss normal polarity chronns have been observed in Chinese loess deposits. The Jaramillo and Olduvai subchrons are well documented, too. On other hand the Blake and the Réunion subchron have been recognised partly. Further short geomagnetic polarity events during the Brunhes chron such as Baffin Bay, Biwa I-III, Fram Strait, Lake Mungo,
Laschamp, Mono Lake, Norwegian-Greenland Sea have not been detected in Chinese loess sediments.

The first complete magnetostratigraphy of the Chinese loess sediments was obtained from the loess/palaeosol sequence in Luochuan, Shaanxi Province (Heller and Liu, 1982, 1984; Torii et al., 1984 see Figure 2.4). The Matuyama/Brunhes boundary (MBB) was found in palaeosol S8 at 53.05 m depth. The Jaramillo subchron was observed between 67.30 m (L10) and 72.50 (S11) and the Olduvai subchron between 107.40 m and 117.40 m (Heller and Liu, 1982). The Gauss/Matuyama boundary (2.6 Ma) occurs at the loess/red clay transition and coincides with the beginning of loess sedimentation in Luochuan.

![Figure 2.5: Correlation of loess/palaeosol sequences over the CLP, after Liu et al., 1991.](image)

Further investigations broadly confirmed the observations from Luochuan. Meanwhile there is overwhelming evidence that the MBB on the CLP is recorded mostly in loess layer L8 (Liu et al., 1988; Cao et al., 1988; Rolph et al., 1989; Rutter et al., 1991; Zheng et al., 1992; Li et al., 1997; Ding et al., 1998; Zhu et al., 1998; Spassov et al., 2001). The most recent magnetostratigraphic
Chinese loess studies locate the geomagnetic reversal boundaries in the following lithologic horizons: The MBB in L8, the Jaramillo subchron between S10 and S12, the Olduvai subchron between L25 and S26 and the Gauss/Matuyama boundary (GMB) in L33.

The formation underlying the loess is the so-called red clay, which contains the chrons Gauss, Gilbert and 3A, and also several subchrons (Sun et al., 1998, Ding et al., 1999, see Figure 2.5). The red clay is also an aeolian sediment, but strongly altered by pedogenesis (Ding et al., 1999). Magnetostratigraphic investigations of the red clay part at Lingtai postulate a basal age of this section of about 7 Ma. The actual dust accumulation and desertification in China begins much earlier at about 22 Ma (Sun et al. 1998; Ding et al., 1999; Guo et al., 2002b).

### 2.2.5 The relationship between loess deposition and climate

Differential heating of the Earth’s surface by the sun leads to the formation of thermal convection cells (Hadley cell). Regarding a non-rotating Earth, cold air at the equator is heated (pressure decreases), rising up and moving aloft toward the poles. There it is cooled (pressure increases), sinking down and moving back to the equator near the surface. Because of the rotation of the Earth the Coriolis force confines the Hadley cells to the tropical belt. The warm air is cooled at the 30° of latitude and sinks down there. Hence, the global circulation conditions are ideal at latitudes around 30° for forming high-pressure systems (anticyclones). A circumpolar band of low pressure occurs at 60° of latitude and the surface winds between 30° and 60° latitude prevail westerly on both hemispheres.

However, the development of anticyclones depends also on the Earth’s surface. Ideal conditions are for instance constant temperature and humidity. North-eastern Asia offers such conditions. The terrain is wide, uniform and without large water bodies. From December to March southern Siberia and Mongolia becomes the site of an intense high-pressure system centred at 100° E and 50° N (Crutcher and Meserve, 1970; cf. Figure 2.6). It produces strong and persistent surface winds, which diverge spiralling from the centre. However, the surface air-flow (0 to 1 km height) in the source regions is controlled by the Siberian anticyclone and the globally westerly winds. Hence, the main winds, which
Figure 2.6: Long-term pressure and temperature distribution over Asia during January and July. The large high pressure system during winter time is responsible for present day loess accumulation (Suchy, 1983). The black ellipse indicates position of the Chinese Loess Plateau and the arrow the main wind direction.
accumulate the loess during wintertime, are blowing from north to north-west (Figure 2.4). In addition, the dry air over the source regions produces large temperature differences and provides ideal conditions for silt production. During summertime the situation is reversed and low-pressure systems mainly form over Asia. Temperature differences are produced between the Asian continent and the Indian/Pacific ocean, because the land surface is faster heated than the water surface. As a consequence a low-pressure cyclone forms over south-west China and the local circulation (sea breeze) brings wet tropical air to Asia causing intensive rain. The long-term temperature and pressure conditions over Asia during the boreal winter and summer are illustrated in Figure 2.6. The annual interplay between summer and winter monsoon is forced to one side during a glacial or interglacial period. The winter monsoon is forced during glacial periods and the intensity of the summer monsoon is reduced. Much more loess is deposited whereas pedogenesis is suppressed. During interglacial periods the summer monsoon is strengthened and creates more humid climate favourable for pedogenesis. Loess accumulation is still going on, but less dust is being transported by the weaker winter monsoon onto the CLP.

2.2.6 Palaeoclimatic significance of magnetic susceptibility

The aeolian dust on the Chinese Loess Plateau does not build up a uniform sediment cover. It contains a large number of embedded palaeosol horizons (Liu and Chang, 1964). Early rock magnetic investigations showed that the magnetic susceptibility also reflects the alternation of loess and palaeosol layers (Li et al., 1974; An et al., 1977). Heller and Liu (1984) quantified the susceptibility differences at Luochuan (Figure 2.4) between loess and palaeosol pointing out that the susceptibility of palaeosol is about twice that of loess. They argued that the process of soil formation under more warm and humid conditions enhanced the susceptibility in contrast to the loess, which has been deposited during arid and cold periods. Further investigations of nearby outcrops (Liu et al., 1985; Kukla and An, 1989) and comparison with sections at Xifeng (Liu et al., 1987; Kukla and An, 1989) and Baoji (Rutter et al., 1991) confirmed these results and interpretations. Loess susceptibility records were successfully correlated with independent climate recorder such as oxygen isotope variations obtained from calcium carbonate shells of marine organisms (Heller and Liu, 1986; Heller and
Evans, 1995). Low values of the oxygen isotope values are well correlated with enhanced susceptibility values (Heller and Liu, 1986).

Several models have been proposed to explain the susceptibility enhancement in the palaeosol. Heller and Liu (1984) explained the increase in magnetic susceptibility by increasing concentration of magnetic minerals due to decalcification and soil compaction. Kukla et al. (1988) assumed a detrital flux of ultrafine magnetite to be responsible for the increased bulk susceptibility. Reduced silt deposition during warm periods concentrates the magnetic particles and enhances susceptibility values. During cold periods the constant detrital input of magnetite from the source regions is then diluted causing lower susceptibility values. Completely different enhancement models such as frequent natural fires (Kletetschka and Banerjee, 1995) have also been proposed. The most widely accepted model considers weathering processes to be responsible for the enhancement of magnetic susceptibility in palaeosols (Zhou et al., 1990; Maher and Thompson, 1991). The role of iron oxide producing bacteria during pedogenesis has been considered, too (Evans and Heller, 1994; Maher, 1988), although they could not positively be identified. Indirect evidence for iron oxide producing bacteria in recent luvisols from southern Germany was given by Hanesch and Petersen (1999).

The succession of glacial and interglacial periods during the Quaternary is strongly controlled by the parameters of the Earth’s orbit. Periodic variations of the oxygen isotope signal in deep-sea sediments have been associated with the frequency content of the insolation curve. The known orbital periods are ~400 and ~100 kyr due to the eccentricity, ~41 kyr due to the obliquity (changes of the tilt angle with respect to the orbital plane) and precession ~19 and ~23 kyr. Spectral analysis has been also applied to susceptibility profiles of loess/palaeosol sequences. Heller and Liu (1986) found only one peak near 40 kyr by matching susceptibility peaks of the Luochuan section to the theoretical insolation curve (also called orbital tuning) of Berger (1978). Wang et al. (1990) determined susceptibility peaks at 40 kyr and 100 kyr in the Baoji section. Lu et al. (1999) re-investigated the Luochuan section and found also non-orbital scale cycles at 200 and 300 kyr beside all the orbital frequencies. Most recently Heslop et al. (2000) developed a new timescale for the Chinese loess by orbital tuning to the insolation curve. Their cross-spectral analysis showed periods of 19, 23 and 41 kyr. Hence, it is confirmed that susceptibility variations of the Chinese loess are driven to a large extent by the motions of the Earth’s orbit.
3. Characterisation of the loess/palaeosol section at Lingtai, central CLP

The central CLP contains the thickest loess deposits in China. The loess in the central sub-region of the CLP mostly forms a continuous, up to 200 m thick cover, also called Yuan (= high table land consisting of thick loess). The Lingtai outcrop is located about 15 km south of the town of Lingtai (Gansu province) at 34.98° N, 107.56° E and ~1340 m above sea level. The Pleistocene loess/palaeosol sequence is about 180 m thick and overlies an aeolian Mio-/Pliocene red clay (130 m) resting on a Mesozoic sandstone basement. The basal age of the loess-red clay sequence is about ~7.0 Ma according to the magnetostratigraphic study of Ding et al. (1999). The present day climate at Lingtai is typical for the central loess plateau - warm and humid with an annual mean temperature and precipitation of about 13 °C and 650 mm/yr (Ding et al., 1998). A total of 33 palaeosol horizons developed sequentially during the Pleistocene representing warm/humid intervals alternating with cold/arid periods during which loess was deposited but remained largely unaltered. The section was sampled continuously by cutting oriented vertical block samples of 30 cm height. Small subsamples ~(2x2x2 cm) were cut from these blocks for palaeo- and rock magnetic analyses resulting in a total amount of 13, 400 cubes. Prof. Yue Leping and his co-workers in China organised collection and preparation of this enormous amount of samples. The Lingtai section shows the typical central CLP loess/palaeosol succession (Liu et al., 1985) and has been well described by Ding et al. (1999). Following numbering system has been established for Chines loess strata. Palaeosols are labelled with letter S and loesses with L, followed by the consecutive number of the palaeosol or loess layer. The first layer is the Holocene soil labelled with S0, the underlying first loess with L1. The Holocene soil (S0) has a Bw horizon that is characterised by moderate organic matter content, complete carbonate leaching and a coarse prismatic structure. However, S0 cannot be characterised easily since the soil structure is destroyed due to intensive farming. Palaeosol S1 only has a blackish brown colour indicating a high organic matter content whereas palaeosols S2 to S14 are rather brownish or reddish. In contrast to palaeosols S15 - S32, carbonate concretion horizons are lacking in the lower B-horizon of S2 to S14. S5 is the thickest palaeosol in the whole section and shows large clay skins in the B-horizon, a hint for distinctive clay translocation. The lithostratigraphic marker horizons S5, L9, L15 are well recognised. The loess layers L9 and L15 containing coarser silt are also well
Figure 3.1a: Litho- and magnetostratigraphy of the Lingtai section. Left: susceptibility and reversal boundaries from Ding et al. (1999), middle: susceptibility and reversal boundaries from this work, right: frequency dependence from this work. The letters B, M, J and O denote Brunhes, Matuyama, Jaramillo and Olduvai geomagnetic polarity chron and subchrons, respectively. The patterned bar at the upper Olduvai boundary indicates an uncertainty as many multiple polarity flips occur over a depth interval of 2.88 m. The distance between both outcrops is ~1.5 km.
Figure 3.1b: Continuation of the record from Figure 3.1a. The letters M and G denote the Matuyama and Gauss chrons.
pronounced at Lingtai. A complete lithologic description (after Ding et al., 1999) is given at http://www.ngdc.noaa.gov/paleo/datalist.html#loess.

Magnetic susceptibility was measured in order to characterise the general lithostratigraphy of the Lingtai section. The susceptibility values of the palaeosols have average values of about 200x10^{-8} m^3/kg in the upper part of the section down to S5 (Figure 3.1a) and contrast the low susceptibility of the loess layers. Loess layer L4 has the lowest susceptibility of the whole profile (~23x10^{-8} m^3/kg). From S5 to L9 the difference between palaeosol susceptibility (~75x10^{-8} m^3/kg) and loess susceptibility (~30x10^{-8} m^3/kg) is smaller. From S9 to S14 the palaeosols have again slightly higher values (~150x10^{-8} m^3/kg). Only exception is L10 with high susceptibilities (~75x10^{-8} m^3/kg). The loess layers below S15 cannot easily be recognised and have sometimes higher susceptibilities than the palaeosols (e.g. L25). The palaeosols from S15 to S24 have low values and from S27 to S32 the values are increasing. Carbonate concretion layers occur in some of the loess horizons (e.g. in L4, L6, L9, L15, L26, L27) with susceptibilities of 7 to 18x10^{-8} m^3/kg. Below L33, the Pliocene red clay starts with rather constant susceptibility values. The F-Factor follows in general the susceptibility curve. Low values characterise loesses and high values palaeosols. Beside many noisy fluctuations, the F-factor oscillates around 10 % over the whole profile. Clearly low F-factors between 1 and 3% are found in loess layers L3, L4, L9, L15, L27, L32, L33 and in the carbonate concretion layers. The F-factor in the red clay is fairly constant with 10% and does not show low values as in the loesses.

The chronology of the Lingtai loess/palaeosol section was determined using magneto-stratigraphy. Based on the preliminary research of Ding et al. (1998) 520 oriented samples have been selected at intervals where reversal boundaries were expected. Almost all samples exhibit an initial normal polarity of the natural remanent magnetisation (NRM) indicating that they carry a present-day secondary overprint. Therefore, the samples have been stepwise thermally demagnetised from 100 to 520 °C whereby the overprint component was removed at temperatures between 250 °C and 300 °C. The characteristic direction of the NRM has been determined using principal component analysis (Kirshvink, 1980). The Matuyama-Brunhes boundary (MBB) has been found in the lower part of loess L8 at ~ 61.75 m. The upper boundary of the Jaramillo subchron has been observed in the upper part of palaeosol S10 at 77.76 m depth.
and the lower boundary at the bottom of S11 at ~83.34 m (Figure 3.2). The upper boundary of the Olduvai subchron is complicated by many polarity flips which occur between 131.08 and 133.92 m depth within loess L25, whereas the lower Olduvai boundary is better defined at 141.6 m in the upper part of S26. The Gauss-Matuyama boundary is found at a depth of ~179 m within the lowest loess layer L33. Nineteen additional samples spread over palaeosol S1 (from 8.42 m to 10.56 m) have been investigated in order to find evidence of the Blake

**Upper Jaramillo, Lingtai**

**Lower Jaramillo, Lingtai**

**Figure 3.2:** Stratigraphic plot of susceptibility (left), VGP (middle), and F-factor (right), for both Jaramillo transitions at Lingtai. Whilst the rock magnetic parameters are constant at the upper Jaramillo transition, they fluctuate at the lower transition. These fluctuations coincide with multiple polarity changes. A similar observation was made by Guo et al. (2002a).
subchron but the samples gave no indication for reversed directions of the characteristic remanence (ChRM).

The susceptibility data and the magnetic reversal boundaries were compared with results of Ding et al. (1998, 1999) who studied a section located not far from ours at 34°58′22″N and 107°33′22″E. The distance between both outcrops is ~1.5 km only (Ding et al., 1999; Figure 2). The susceptibility data of Ding et al. (1999) generally agree with our measurements (Figure 3.1a/b). Minor differences are recognised for instance in S1 (at ~9 m) and S7 (at ~55 m), which have two peaks in Ding’s et al. (1999) outcrop but only one in ours. Also the high susceptibility plateau within L1 (corresponding to oxygen isotope stage 3) is much more pronounced in our outcrop than in Ding’s et al. (1999). The comparison shows further, that the stratigraphic layers seem to be displaced between the two sections towards lower depths in our Lingtai outcrop. The top of palaeosol S2 occurs at 18.1 m in the Ding et al. (1999) description but at 19.1 m in ours. The depth difference increases with depth being 4.8 m at S8, 7.16 m at S14, 7.6 m at S26 and 8.6 m S32. The maximal displacement at S32 is quite high with 8.6 m. We attribute the continual displacement to local sedimentation differences or systematic levelling errors in the steep cliffs of the loess plateau valleys.

The magnetic reversal boundaries are displaced, too. They were however found in both outcrops in the same stratigraphic units, often not at the same stratigraphic level (Figure 3a/b). For instance, the lower boundary of the Olduvai subchron is placed in the middle of S26 by Ding’s et al. (1999) data, whereas our study puts it at the top of S26.
Part II

Research
The Matuyama/Brunhes geomagnetic polarity transition at Lingtai and Baoji, Chinese Loess Plateau

Summary

Two sections on the Chinese Loess Plateau (CLP) at Lingtai and Baoji have been investigated to obtain detailed information about the exact stratigraphic position and the recording quality of the Matuyama/Brunhes geomagnetic polarity boundary (MBB). The continuously sampled MBB occurs in the glacial period loess layer L8 in both sections at a similar profile depth indicating comparable average sedimentation rates of about 8 cm/kyr. This stratigraphic position has been recognised in many previous Chinese loess studies and contrasts with observations in marine sediments where this polarity transition is found at the beginning of interglacial oxygen isotope stage 19. The directional patterns of the characteristic remanent magnetisation (ChRM) of the two transitional loess records are inconsistent and even depend on the demagnetisation technique used to isolate the ChRM. The MBB records also differ significantly from those obtained from the loess section at Weinan and at ODP site 792A near Japan. The virtual geomagnetic pole (VGP) paths recorded in loess are not well defined, switch polarity several times and do not prefer specific longitudinal bands. It is suggested that the loess MBB transitional records do not reflect geomagnetic field variations directly, but are the result of a complicated magnetisation lock-in mechanism. By comparison with the astronomically-tuned prediction, this process has been delayed in the loess sediment column by 1.5 to 2 m corresponding to about 25'000 yr after deposition.
4.1 Introduction

Despite many years of investigation, the actual process by which in particular Chinese loess has been magnetised, is still poorly understood. The major normal and reversed polarity chronos of the last 2.5 Myr have been recognised, but there is a serious conflict over the position of the Matuyama-Brunhes Boundary (MBB) in marine and continental records (Hus and Han, 1991; Tauxe et al., 1996; Zhou and Shackleton, 1999; Heslop et al., 2000). The observation of the MBB (and other polarity boundaries) is of utmost stratigraphic importance since other dating methods can not be applied or have failed. As summarised by Tauxe et al. (1996), the MBB is placed in marine sediment sections in the lower part of oxygen isotope stage 19, which corresponds to an interglacial stage. In most Chinese loess sections, however, this polarity transition is observed in the middle or lower part of loess layer L8 (e.g. Heller and Evans, 1995). In an attempt to shed further light on this problem, the MBB of two loess sections from Lingtai and Baoji, which have similar mean sedimentation rates, has been studied in detail and will be compared with another loess section studied by Zhu et al. (1994) at Weinan (34.2°N, 109.2°E) on the Chinese Loess Plateau and with ODP core 792A (31.0°N, 140.0°E) off the south-east coast of Japan (Cisowski et al., 1992).

4.2 Geological setting, sampling and rock magnetic properties

The Lingtai loess-palaeosol section is located about 15 km to the south of the town of Lingtai, Gansu Province (34.98°N, 107.56°E) at an elevation of 1340 m above sea level. The section is about 175 m thick and overlies Mio-/Pliocene aeolian red clays, which rest on a Mesozoic sandstone basement. A total of 33 soil horizons developed sequentially during the Pleistocene representing warm/humid intervals alternating with cold/arid periods during which loess was deposited (Ding et al., 1999). A first set of oriented block samples (set A) was collected to provide continuous coverage of the section for about 3 m from palaeosol S8 through the slightly weathered loess L8 into palaeosol S7. The MBB was observed at a depth of approximately 62 m in the middle part of loess layer L8. This indicates an average accumulation rate of ~8 cm/kyr. Each of the
150 samples (2x2x2 cm) thus represents an average time interval of 250 yr. All samples of set A were demagnetised thermally. Another independent set of oriented block samples (set Y, collected about 1 m away from set A) was demagnetised using alternating fields. A secondary present-day component of the natural remanent magnetisation (NRM) was removed at temperatures between 250 °C and 300 °C or at peak fields of 15 mT or 20 mT for loess and palaeosol, respectively (Fig. 4.1). The characteristic remanent magnetisation (ChRM) of the samples plotted in Fig. 4.1 is reversed and corresponds to the Matuyama chron. Isothermal remanent magnetisation (IRM) and anhysteretic remanent magnetisation (ARM) are much stronger, and coercivity spectra are much narrower, in S8 than in L8 suggesting prominent secondary (bio)chemical formation of ferrimagnetic material in the palaeosol.

Specific low field susceptibility, coercivity, coercivity of remanence, saturation magnetisation and saturation remanence vary smoothly with stratigraphic position and do not show any abrupt changes where the MBB occurs (Figs. 4.2 and 4.3). Coercivity values, slightly increasing from S8 to L8, are in agreement with the ARM and IRM results and suggest that two coercivity populations, one of detrital and the other of in situ origin coexist in the loess/palaeosol samples (cf. Evans and Heller, 1994). Their contributions depend on lithology and hence on palaeoclimate. Higher concentration of a superparamagnetic mineral fraction formed in situ is indicated in palaeosol S8 by higher F-factors and saturation magnetisations together with low coercivities.

The section at Baoji (Shaanxi Province) is located 5 km north of the city of Baoji (34.4°N, 107.1°E) at an elevation of approximately 970 m, about 120 km south of Lingtai. The section is 159 m thick and contains in total of 37 identifiable palaeosols (Rutter et al., 1991). After thermal treatment, the MBB was found at a depth of 58.5 m, again in loess layer L8 as in Lingtai (Fig. 4.3).

4.3 MBB directional change

Concerning the VGP latitude profiles in Lingtai (Fig. 4.3), seven full polarity changes are indicated during the transitional interval, but they occur at slightly different profile depths in the two data sets. The thickness of the transitional
demagnetisation treatment in both lithologies. A present-day viscous overprint is removed by about 20 mT or 250 °C demagnetisation techniques yield clearly defined reversed characteristic magnetisation representative samples from loess layer L8 and underlying palaeosol S8 from Lingtai. Both magnetisation (IRM), anhysteretic remanent magnetisation (ARM) and hysteresis properties of the mineral fractions present.

Figure 4.1: NRM thermal and alternating field demagnetisation (AF), isothermal remanent magnetisation (IRM), anhysteretic remanent magnetisation (ARM) and hysteresis properties of representative samples from loess layer L8 and underlying palaeosol S8 from Lingtai. Both demagnetisation techniques yield clearly defined reversed characteristic magnetisation directions. A present-day viscous overprint is removed by about 20 mT or 250 °C demagnetisation treatment in both lithologies.

VGP zone amounts to about 0.4 m in both sets. Two intermediate VGP positions are observed in the set Y data within 0.5 m below the first full reversal. They are not present in the data set A. These differences reflect different responses to the demagnetisation treatment of the coercivity and blocking temperature distributions of the mineral fractions present.

In order to assess the reliability of the data, the maximum angular deviation (MAD) of the regression line of the ChRM was calculated, using for all samples the same number of treatment steps. Both demagnetisation methods yield approximately the same precision above and below the transition zone with MADs generally below 10° whereas the ChRM directions are not well defined within this zone, with MADs sometimes exceeding 30°. The two intermediate VGPs mentioned above also have high MAD values. The parameters NRM$_{20mT}$/χ and NRM$_{20mT}$/M$_{rs}$ are often taken as relative palaeointensity
indicators in sediments. Their sequence can be divided into an upper part where the signal is fairly constant, a middle part with very low values within the transition zone and a lower part where these ratios show appreciable variation. The low values in the mixed polarity interval are interpreted to be caused by the simultaneous occurrence of grains carrying reversed and normal ChRM. The very weak ChRM intensities are produced by the competing magnetisations rather than by a low intensity geomagnetic field (see also Rutter et al., 1991). It is not clear at this stage if the variations of both parameters above and below the transition zone reflect geomagnetic field behaviour.

At Baoji, the transitional VGP interval spans about 0.5 m and displays five full polarity changes, and some intermediate directions which are not observed.

Figure 4.2: Low field susceptibility (\(\chi\)), frequency dependence of susceptibility (F-factor resulting from measurements at two frequencies), coercivity \(H_c\) (dotted), coercivity of remanence \(H_{cr}\) (solid), saturation magnetisation \(M_s\) (not completely saturated in the maximum field of 300 mT; dotted) and saturation remanence \(M_{rs}\) (solid) and the ratios of these parameters. All data are from set Y in Lingtai, except the dotted susceptibility curve, which is for set A. Since both susceptibility curves coincide, stratigraphic consistency of both sample sets is demonstrated.
at Lingtai (Fig. 4.3). The susceptibility decreases steadily from S8 reaching relatively low values in L8.

The data from Lingtai and Baoji can also be compared with the loess-palaeosol section at Weinan (Zhu et al., 1994). This section is located about 160 km ESE from Lingtai and consists of a sequence of 33 loess-palaeosols with a total thickness of about 150 m. Using thermal demagnetisation, the Matuyama-Brunhes transition was observed in loess layer L8 which accumulated at a rate of about 9.3 cm/kyr. Three full polarity changes are recognised here over a transition interval of about 0.5 m. The polarity transitions, however, contain a substantial number of intermediate VGPs especially during the last apparent polarity change and are much less abrupt than those at Lingtai and Baoji (Fig. 4.4). The CLP data can be further compared with ODP site 792A – the nearest marine record – which has a similar sedimentation rate (about 8.1 cm/kyr, Cisowski and Koyama, 1992). This marine MBB record shows one sharp polarity transition spanning < 5 cm, like many other marine MBB records (e.g. Clement and Kent, 1987).

Figure 4.3: Directional behaviour of the MBB transitional polarity interval in the sections at Baoji (left) and Lingtai (right). The parameters NRM$_{20mT}$/M$_{rs}$ (dotted) and NRM$_{20mT}/\chi$ (solid line) yield relative palaeointensity results, which vary consistently throughout the profile. MAD values have been calculated using always five demagnetisation steps. At Lingtai, seven full polarity changes are observed in both VGP data sets. A slight stratigraphic shift of the transitional directions is seen between thermally and AF demagnetised data. At Baoji, five apparent full polarity changes characterise the MBB transition.

The data from Lingtai and Baoji can also be compared with the loess-palaeosol section at Weinan (Zhu et al., 1994). This section is located about 160 km ESE from Lingtai and consists of a sequence of 33 loess-palaeosols with a total thickness of about 150 m. Using thermal demagnetisation, the Matuyama-Brunhes transition was observed in loess layer L8 which accumulated at a rate of about 9.3 cm/kyr. Three full polarity changes are recognised here over a transition interval of about 0.5 m. The polarity transitions, however, contain a substantial number of intermediate VGPs especially during the last apparent polarity change and are much less abrupt than those at Lingtai and Baoji (Fig. 4.4). The CLP data can be further compared with ODP site 792A – the nearest marine record – which has a similar sedimentation rate (about 8.1 cm/kyr, Cisowski and Koyama, 1992). This marine MBB record shows one sharp polarity transition spanning < 5 cm, like many other marine MBB records (e.g. Clement and Kent, 1987).
The VGP path at Weinan shows clearly a large number of intermediate VGPs which occupy a broad longitudinal band across Australia and eastern Asia (Fig. 4.5) whereas the ODP 792A VGP path runs across the Americas as do the precursory excursions ~50 cm below the main reversal at 58.2 m. The thermally cleaned data from Baoji and Lingtai jump from the reversed pole position to the normal without any intermediate values. The AF cleaned Lingtai data contain a few intermediate poles in the western Pacific. Thus the transition records are characterised by a high inconsistency with respect to the frequency of apparent reversals within the MBB reversal, their stratigraphic position as well as the VGP tracks themselves.

**Figure 4.4:** Profiles of VGP latitudes spanning the Matuyama-Brunhes boundary of three CLP sites and the nearby ODP 792A site. The thickness of the transitional zones in the loess sections is similar but the number of complete polarity changes varies (seven at Lingtai, five at Baoji, three at Weinan). One polarity change spanning < 5 cm is observed at ODP site 792A. The Weinan profile has been plotted on a relative depth scale only.
4.4 Discussion

The co-latitudinal sites discussed here are separated in longitude by about 30° only. Therefore their VGP paths during polarity transitions are expected to be very similar (McFadden et al., 1993) and should fall within a 60° longitudinal sector centred either at 120°E or 60°W (Tric et al., 1991; Laj et al., 1992). For a R to N transition observed at east Asian sites, symmetry considerations predict VGP paths along the eastern segment (Gubbins and Love, 1998). Only the Weinan data seem to comply with these expectations. The OPD792A path is far off this sector and the precursory intermediate VGPs in the Lingtai AF profile fall between the two. Furthermore, it is important to recall that Valet et al.

Figure 4.5: VGP paths of the five MBB records plotted in Figure 4.4. The paths are not clearly defined and cross the equator on different tracks. This inconsistency is considered incompatible with a geomagnetic origin of the VGP paths in the loess sections.
(1992) have vigorously contested the notion of longitudinal confinement of VGP paths. In addition to the disagreement of the VGP paths with geomagnetic reversal models, the question of the process of remanence acquisition in loess sediments must be addressed. This process depends on environmental conditions during aeolian deposition and eventual subsequent pedogenesis of variable type and degree. A simple depositional remanence (DRM) is to be excluded following the arguments of Hus and Han (1991) and Zhou and Shackleton (1999) who document significant MBB displacement in loess with respect to the Australo-Pacific meteoritic glass layer observed in marine sediments. A post-depositional remanence (pDRM) in pelagic sediments is apparently delayed by at most a few centimetres (Tauxe et al., 1996), but the boundary in Baoji and Lingtai has been locked 1.5 - 2 m below the astronomically expected occurrence level (Heslop et al., 2000). Translating this offset into time using the estimated average sedimentation rates, the MBB is delayed by about 15000 to 25000 years.

Detrital magnetisation, however, may not be the only magnetisation process. The loess in the central, southern and eastern CLP is generally not pristine compared to the unaltered loess in the western CLP (Evans and Heller, 1994). Not only palaeosols, but also loess samples from these regions carry a substantial amount of viscous remanent magnetisation (VRM) which is much less or nearly absent in the western CLP. The VRM resides in a secondary, low coercivity, mineral of (bio)chemical origin formed in situ. At Lingtai, $H_c$ and $H_c$ steadily increase from S8 to L8 whereas $M_s$ and $M_{rs}$ decrease simultaneously and at the same rate ($M_{rs}/M_s$ constant). The secondary material is obviously less abundant in the upper part of L8, but still exists throughout this layer (Fig. 4.2). Nevertheless, it remains unclear at present whether the NRM was acquired as pDRM or delayed CRM.

Recently, Bleil and v. Dobeneck (1999) showed that polarity events in marine calcareous oozes are recorded only if they last longer than half the lock-in depth divided by the sedimentation rate. If this calculation – based on a simple pDRM model – applies also to loess sediments on the CLP, records of short polarity events such as the Blake event and others may be significantly disturbed or not observed at all.
4.5 Conclusions

1. It is suggested that the inconsistent VGP paths of the loess/palaeosol records do not represent details of the MBB transitional geomagnetic field behaviour. The extremely low NRM$_{20mT}/\chi$ and NRM$_{20mT}/M_r$ ratios, the high MAD values, the variable ChRM polarity patterns obtained using different demagnetisation techniques and the multiple polarity changes in the loess MBB transition zones are interpreted as reflecting complex magnetisation lock-in processes. They appear to incorporate variable detrital and chemical contributions, both in palaeosols and in loess layers. Assuming a constant sedimentation rate, these processes have been delayed by up to 25000 yr after deposition as was inferred from careful stratigraphic correlation (Heller et al., 1987; Zhou and Shackleton, 1999) and astronomical tuning (Heslop et al., 2000).

2. Loess from most of the Chinese Loess Plateau does not seem to be a high fidelity recorder for geomagnetic field transitions probably because of a very complex magnetisation process. Investigations on the western Chinese Loess Plateau may be more promising because low weathering state and higher sedimentation rates offer simpler recording conditions.

3. The east Asian VGP reversal paths of the Matuyama-Brunhes boundary observed in different environments (continental loess and marine volcaniclastics) are not in agreement. Comparison of the VGP data under consideration (except data from Weinan) with proposed phenomenological reversal models is inconclusive.
Detrital and pedogenic magnetic mineral phases in the loess/palaeosol sequence at Lingtai (Central Chinese Loess Plateau)
Abstract

A detailed rock magnetic investigation of loess/palaeosol samples from the section at Lingtai on the central Chinese Loess Plateau (CLP) is presented. Bulk and grain size fractionated samples have been analysed using coercivity spectra of remanence acquisition/demagnetisation curves, which identify four main remanence carriers in different grain size fractions of loesses and palaeosols. Thermal demagnetisation of isothermal remanent magnetisation (IRM) and Curie temperature measurements identify magnetite, maghemite of different oxidation state and hematite as remanence carriers. A linear source mixing model quantifies the contribution of the four components. Up to two thirds of the total IRM of the palaeosols are due to slightly oxidised pedogenic magnetite. Two detrital components dominate up to 90 % of the IRM of the loess samples and were identified as maghemite of different oxidation degree. Detrital hematite is present in all samples and contributes up to 10 %. The iron content of the grain size fractions gives evidence that iron in pedogenically grown remanence carriers does not originate from the detrital iron oxides, but rather from iron-bearing clays and mafic silicates. The contribution of pedogenic magnetite to the bulk IRM increases with increasing pedogenesis, which depends in turn on climate change.
5.1 Introduction

From deposition to the final stages of diagenesis, sediments are affected by many environmental processes. Transport processes provide detrital input and weathering processes cause dissolution of minerals and iron release for the inorganic or biochemical neoformation of magnetic minerals. These processes are mainly driven by climate, as it has been demonstrated for the loess/palaeosol sequences of the Chinese Loess Plateau (CLP) (e.g. Heller and Evans, 1995). Over the past 2.6 Myr, loess has been deposited during cold and arid periods whereas palaeosols developed under warm and humid conditions. Magnetic minerals have been enriched during soil formation and the magnetic susceptibility of palaeosols has been enhanced. The susceptibility time series of the Chinese loess is dominated by the Earth’s orbital parameters. Precession, obliquity and eccentricity periods have been recognised (Heller and Liu, 1986; Wang et al., 1990; Heslop et al., 2000), hence proving the close assignment between physical rock properties (e.g. susceptibility and grain size) and climate.

Magnetic and geochemical parameters have been used for quantitative estimates of climate parameters such as palaeoprecipitation. The $^{10}$Be flux is considered to consist of a constant atmospheric component and a variable dust component, which records high $^{10}$Be fluxes during glacial times (Shen et al., 1992). The magnetic susceptibility is also composed of two components: a dust component (detrital) and an in-situ formed component (pedogenic). If both dust fluxes are assumed to be proportional, then the pedogenic susceptibility flux can be extracted from the total signal (Beer et al., 1993). Heller et al. (1993) estimated the pedogenic susceptibility component and modelled the palaeoprecipitation for the last 130 kyr at Luochuan (central CLP). The pedogenic susceptibility flux was converted into palaeoprecipitation by linear correlation of present day precipitation and modern soil susceptibility from the CLP. Maher and Thompson (1995) re-constructed palaeoprecipitation using a logarithmic relation between present day precipitation and soil susceptibility. Their pedogenic susceptibility component was estimated simply from the difference between palaeosol susceptibility and that of the most silty and unweathered loess beds.

Magnetic methods offer the great advantage over other methods like X-ray diffraction or colourimetry that extremely low concentrations of ferromagnetic minerals can be detected and identified without costly sample preparation and
within short measurement time. For instance, concentrations of \(~0.12\) ppm (by volume) of pure hematite or goethite and \(~0.79\) ppb (by volume) of pure maghemite or magnetite can be identified using a modern vibrating sample magnetometer. Bulk magnetic measurements represent a smoothed integral over all magnetic mineral components present. Bulk magnetic properties (such as susceptibility, median destructive field, S-ratio, Day-plot parameters etc.), however, may not give clear evidence about the contribution of different magnetic minerals in loess/palaeosol samples.

The amount, type and grain size of the individual magnetic mineral phases present in a rock can be derived from coercivity distributions of acquisition or demagnetisation of laboratory produced magnetisations. Robertson and France (1994) were the first to model acquisition curves of isothermal remanent magnetisation (IRM) in order to quantify magnetic mineral populations from different sources – also called endmembers – present in a sample. This unmixing technique was further developed by Stockhausen (1998), Kruiver et al. (2001) and Heslop et al. (2002). Each individual mineral component is characterised by a single lognormal function. It is assumed that the bulk IRM corresponds to the sum of the contributions of all individual components/endmembers (source mixing model). Carter-Stiglitz et al. (2001) could show indeed that the magnetisations of their materials are mixing linearly. In order to model a measured IRM acquisition curve, an unknown possibly infinite number of lognormal functions is necessary. Therefore source mixing or unmixing can result in complicated calculation problems using the techniques available. Egli (2003) proposes a generalised function for coercivity populations that models endmember distributions without assuming specific (e.g. lognormal) distributions. This approach reduces the number of distributions needed to fit the experimental results.

The method of source mixing can be applied to the loess/palaeosol sediments of the CLP. Evans and Heller (1994) proposed the coexistence of two magnetic mineral components in loess/palaeosol sediments on the CLP and tried to estimate their contributions from the derivatives of IRM acquisition curves. The first component is a detrital population of magnetic particles, which is always present and appears to be uniform across the loess plateau. The second component is a superimposed authigenically grown population varying from layer to layer within a section and from site to site. Eyre (1996) extended this model to a four-component model.
Table 5.1: Low field susceptibility, IRM at 4.5 T and total iron and titanium content of four grain size fractionated loess and palaeosol samples from Lingtai. The total susceptibilities recalculated from the contribution of the grain size fractions agree rather well with the susceptibility of the bulk samples. The small difference in Li-L4 may be due to the decalcification of the grain size fractions.

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<th>Name</th>
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<th>Total iron</th>
<th>Total titanium</th>
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The present study takes a different approach in identifying remanence carriers and their grain size. We measured IRM properties of different bulk and of grain size fractionated loess/palaeosol samples and calculated their coercivity spectra. The coercivity spectra of the fractionated samples were used to evaluate the endmembers present in a bulk sample. IRM curves of bulk samples were then modelled and reconstructed by varying the contribution of the modelled endmembers. Using this approach, we seek to obtain more information about the interaction between climate, deposition, alteration and pedogenesis on the central CLP.

5.2 Sample description

The samples have been collected from the loess/palaeosol sequence at Lingtai (34.98 °N, 107.56 °E), Gansu Province. The upper part of this section consists of 33 pairs of alternating loess/palaeosol layers down to 175 m profile depth. The part below is formed by Mio-/Pliocene red clay with a basal age of about 7.05 Ma (Ding et al., 1998) at a depth of 305 m.

In order to cover a wide range of magnetic properties, samples have been selected from the well developed palaeosols S1 and S5 and from the pristine loess L4. Weakly developed soils (S7, S8) and loesses of different weathering stage (L7, L8) have been considered, too. Palaeosol S7 is genetically related to loess layer L8, because it formed partly by alteration of loess L8 during warmer climate conditions. An additional loess sample has been chosen from the western CLP where pedogenesis is weaker in a much more arid climate than on the central CLP. This sample (By-L4) originates from loess layer L4 of the Baicaoyuan section (35.7 °N, 104.9 °E). Evans and Heller (1994) considered this sample to be not affected by pedogenesis (Table 5.1).

5.3 Experiments and procedures

Two types of samples, bulk samples and grain size fractionated samples, were analysed. The samples to be fractionated were disaggregated and decalcified using standard laboratory procedures (Scheinost et al., 2002). The disaggregated samples were wet-sieved (mesh size 50) to separate the fraction > 50 µm. The
remaining split \( \leq 50\, \mu m \) was fractionated by sedimentation in water into grain size classes of 20 - 50 \( \mu m \), 5 - 20 \( \mu m \) and 2 - 5 \( \mu m \). Grains \( \leq 2\, \mu m \) were centrifuged into size fractions of 0.2 - 2 \( \mu m \) (coarse clay) and \( \leq 0.2\, \mu m \) (fine clay). Parts of all six fractions have been mixed with wax (Hoechst Wachs C) and pressed into pellets, with a constant mass ratio of sample/wax (\( = 4.4 \)). The iron and titanium concentrations of the fractionated samples were measured using a X-Lab 2000 energy-dispersive X-ray fluorescence spectrometer (Spectro) equipped with a sequence of secondary targets (Mo, Al\(_2\)O\(_3\), B\(_4\)C/Pd, Co and HOPG) to generate polarised X-rays. The lower detection limit is 0.5 mg/kg (\( = 5\times10^{-5}\% \)).

Low and high coercivity minerals are expected in the loess/palaeosol samples. In order to determine the low coercivity mineral content of the original bulk samples, anhysteretic remanent magnetisation (ARM) and isothermal remanent magnetisation (IRM) was demagnetised by alternating fields. After initial demagnetisation in an AC (alternating current) peak field of 300 mT, the samples were given an isothermal remanent magnetisation in a DC (direct current) field of 300 mT (IRM\(_{300mT}\)). An AC peak field of 300 mT with a superimposed DC field of 0.1 mT was utilised for the ARM experiment. Both laboratory remanences were measured and demagnetised using a 2G cryogenic magnetometer with built-in AF coils. AF demagnetisation along the magnetised axis was started after passing a constant waiting time of 3 min using 45 logarithmically distributed steps between 0 and 300 mT.

The high coercivity mineral content of the fractionated samples was tested by stepwise acquisition of IRM using an impulse magnetiser. The pressed pellets were pulverised again and part of this powder (containing sample and wax) was pressed into small gel cups. These gel cups were then magnetised step by step using 38 logarithmically distributed data points ranging from 1 to 4.5 T and always measured after a waiting time of 3 minutes when the IRM had become stable within measurement time. The magnetisation of the samples has not been corrected for the wax content because of its small mass and remanence contribution.

The remanence acquisition and demagnetisation spectra have been calculated following the method of Egli (2003). His method is summarised here as follows: The field of the measured curve is scaled to obtain a symmetric sigmoidal shape of the graph. A scale transformation of the magnetisation linearises the curve. The scaled graph is then fitted with a hyperbolic tangent function, which is
assumed to be the (unknown) noise-free magnetisation curve. The residuals between measured and fitted curve are calculated and the experimental errors are much enhanced at this step. A Butterworth low-pass filter performs the efficient noise reduction, whereby the appropriate choice of filter order and cutoff frequency is eased by the comparison of the filtered and unfiltered Fourier spectra of the residual curve. The subsequent backward transformation results in a fairly noise-free magnetisation curve. Next, the noise-free magnetisation curve is scaled using a logarithmic field scale (base 10). Thus the field axis becomes unitless and the resulting derivative – also called logarithmic coercivity spectrum (LCS) – has the same units as the magnetisation (see Figure 5.1). The maximum error amplitude of the LCS is estimated by comparing measured and filtered magnetisation curve and is shown as errorband.

Thermal demagnetisation of the IRM completes the magnetic mineral characterisation. Powder samples have been mixed and shaken with CC High Temperature (OMEGA®) cement in the mass-ratio 1:6 under dry conditions.

Figure 5.1: Acquisition curve of isothermal remanent magnetisation and calculated logarithmic IRM gradient (coercivity spectrum) of sample Li-L4, fraction < 0.2 µm. The acquisition curve – diamonds represent the individual measurements – shows two major slopes around 1.41 (~26 mT) and 2.78 (~610 mT) and the derivative of the acquisition curve exhibits two separate maxima. The grey band of the gradient curve represents the experimental errors. The maximal applied field amounts to 4500 mT.
After adding a liquid binder the samples were stirred and dried for 24 h in the shielded room of the laboratory. These samples were also AF demagnetised along three orthogonal axes using a peak field of 300 mT. Three IRM coercivity windows have been chosen to analyse different coercivity populations thermally. The samples were first magnetised at a higher DC field value and then AF demagnetised by a smaller field. The field pairs were: [40 mT, 20 mT]; [120 mT, 70 mT] and [1500 mT, 280 mT] using one specimen per field pair. The remaining IRM was demagnetised thermally after a 48 h waiting time to reduce contribution of viscous remanences. The samples were heated for about 40 minutes after the pre-adjusted temperature within the sample zone of the oven was reached. The susceptibility of the samples was monitored to assess mineralogical changes during the heating process after each of the 32 to 35 demagnetisation steps. Blank samples have been prepared in the same way for each magnetisation window and the magnetisation of the samples was corrected for the cement magnetisation. This method was applied to grain size fractionated samples (field pairs: [40 mT, 20 mT] and [120 mT, 70 mT]) and to bulk samples (field pair: [1500 mT, 280 mT]).

The Curie temperature of some grain size fractionated powder samples was measured using a Curie balance of the Kazan State University. Pressed powder samples were heated in air from –196 °C to 700 °C in fields of 50 mT (palaeosol fractions) and 160 mT (loess fractions). A fast heating rate of 150 °C/min was chosen to reduce chemical alteration during heating. Each sample was heated twice in order to assess mineralogical changes.

The susceptibility of the original bulk samples and of the grain size fractionated samples was measured using a KLY-2 susceptibility bridge. As the grain size fractionated samples were mixed with wax, the wax susceptibility was subtracted. Each measurement was repeated 5 times because of the partly very weak signals.

5.4 Results

The initial ARM intensity (Figure 5.2, upper panel) divides the palaeosols clearly in two categories: weakly (Li-S7, Li-S8) and strongly magnetic palaeosols (Li-S1, Li-S5), respectively. This is also reflected by their susceptibilities (Table 5.1). The common property of these palaeosols,
AF demagnetisation of anhysteretic remanent magnetisation (ARM)

**Palaeosols**

AF demagnetisation of anhysteretic remanent magnetisation (ARM)

**Loesses**

AF demagnetisation of isothermal remanent magnetisation (IRM)

**Palaeosols**

**Loesses**
independent of their initial intensity, is the consistent LCS amplitude peak \( P \) at 22 mT, representing one rock magnetic mineral population only. The initial ARM intensities of the loesses are one order of magnitude lower compared to the palaeosols. The maximal LCS amplitude \( (D_2) \) location decreases from 38 mT in the pristine By-L4 to 29 mT in the more weathered Li-L7. Similar to the palaeosols, one magnetic component with a wider coercivity spectrum predominates the ARM spectra of the loesses.

AF demagnetisation of IRM yields different results and indicates the coexistence of two coercivity populations in palaeosols and loesses (Figure 5.2, lower panel). The loess spectra have a major peak around 50 - 60 mT \( (D_1) \). The long wide tail towards lower fields indicates a second low coercivity population \( (D_2) \) peaking around 28 mT. The palaeosols in contrast exhibit a well defined maximum around 20 mT \( (P) \) and a long tail towards higher fields indicating another component \( (D_1) \) which peaks around 70 mT. The lowest coercivity component \( (~ 20 \text{ mT, component } P) \) as well as the intermediate peak around 28 mT \( (D_2) \) occur in both, IRM and ARM spectra whereas the high coercivity peak \( (~ 70 \text{ mT, } D_1) \) is observed in the IRM spectra only. ARM spectra are generally narrower than IRM spectra.

The IRM coercivity distributions of the grain size fractionated samples (Figure 5.3, Table 5.2) do not only depend on lithology, but also on grain size. The smallest fraction \((< 0.2 \mu m)\) contains a low coercivity component, which peaks between 21 and 27 mT in all samples. A high coercivity component between 560 - 760 mT \( (H) \) is also present in this fraction, preferentially in the loesses. The high coercivity part still contributes to the spectra in the fraction 0.2 - 2 \( \mu m \) in which the low coercivity component moves to slightly higher values. The palaeosols peak now at 31 mT \( (P) \) and the loesses at 40 mT to 45 mT. An
Figure 5.3: Normalised spectra of IRM acquisition curves of six grain size fractions of two loesses (Li-L4, Li-L8) and two palaeosols (Li-S1, Li-S7). Absolute values are given in Table 5.2. Four main components (bold letters) can be recognised in all four samples. Component H between 560 and 760 mT can be recognised best in fraction < 0.2 µm, component P at (31 ± 4) mT in fraction 0.2 - 2 µm, component D1 at (113 ± 13) mT in fraction 5 - 20 µm and component D2 at (79 ± 10) mT in fraction 20 - 50 µm.
<table>
<thead>
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<tr>
<td>&lt;0.2 µm</td>
<td></td>
<td>Peak field amplitude</td>
<td>Maximal amplitude</td>
<td>Error at max. amplitude</td>
</tr>
<tr>
<td></td>
<td>mT</td>
<td>mAm²/kg</td>
<td>%</td>
<td>mT</td>
</tr>
<tr>
<td></td>
<td></td>
<td>26</td>
<td>0.57</td>
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<td></td>
<td></td>
<td><strong>610</strong></td>
<td><strong>0.20</strong></td>
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<tr>
<td></td>
<td></td>
<td>~ 125</td>
<td>1.17</td>
<td>8.48</td>
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<td></td>
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<td>463</td>
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<td>2 - 5 µm</td>
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<td></td>
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<td>5 - 20 µm</td>
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<td>Li-S7</td>
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<tr>
<td>&lt;0.2 µm</td>
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<td>Peak field amplitude</td>
<td>Maximal amplitude</td>
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<td>20 - 50 µm</td>
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<td>3.85</td>
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<td></td>
<td>~ 100</td>
<td>3.02</td>
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intermediate component around 140 mT seems to be present, too. This component is strongly developed in the grain size fraction 2 - 5 µm of the loesses, but is much less prominent in the palaeosols, especially in the well developed palaeosol Li-S1. The palaeosol spectra are dominated by a 34 mT peak which is much reduced in the loesses. The loess grain size fractions 5 - 20 µm exhibit only a high coercivity component at 113 mT with almost identical and skewed spectra. The 5 - 20 µm fraction of Li-S1 is predominated by a low coercivity component, whereas the higher coercivity component dominates in Li-S7. The spectra of both loesses are again identical and symmetrical and peak at 79 mT in grain size fraction 20 - 50 µm. The spectra skewness and the LCS maxima of both fractions in Li-S1 are nearly equal. The spectrum of the fraction 20 - 50 µm of Li-S7 seems to be dominated equally by low and high coercivity components. The spectrum of the largest grain size fraction (> 50 µm) is similar to that of the 5 - 20 µm fraction.

Figure 5.4 shows the IRM acquisition spectra of the fractionated samples on an absolute intensity scale. Amongst the stronger magnetic coarse fractions, fraction 5 - 20 µm has the highest magnetisation in the pristine loess Li-L4, whereas the two smallest fractions (< 0.2 and 0.2 -2 µm) are much weaker. The slightly weathered loess Li-L8 is quite similar, but has the strongest magnetisation in the coarse fraction 20 - 50 µm. Again the smallest fractions are weakly magnetic with the same order of magnitude as in Li-L4. The palaeosols behave completely different. The fine grain size fraction 0.2 - 2 µm is by far the most magnetic one. Its maximal amplitude in Li-S1 is very prominent and five times higher than in Li-S7. Accordingly the other fractions are of subordinate importance in Li-S1. The coarser fractions – in particular fraction 5 - 20 µm – gain in importance in Li-S7. Interestingly, the maximal amplitude of this fraction is equal in the genetically related loess Li-L8. The IRM acquisition spectra of the different grain size fractions demonstrate, that mainly four coercivity components (bold in Table 5.2) carry stable remanences in loesses and palaeosols: P at (31 ± 4) mT, D₁ at (113 ± 13) mT, D₂ at (79 ± 10) mT and H at (610 ± 150) mT.

Table 5.2: Numerical values of IRM components observed in six grain size fractions of loess and palaeosol samples (cf. Figures 5.3, 5.4). Peak field is the field at which a component reaches its maximal amplitude. The underlined components have been used as endmembers in the linear mixing model.
The different grain size fractions apparently represent different coercivity populations. This is confirmed by the thermal demagnetisation of IRM and Curie temperature measurements. Figure 5.5a presents thermal demagnetisation results of the IRM given in the coercivity windows as discussed in section 3. The intensity of the low coercivity component (= remanences from grains with coercivities between 20 and 40 mT) of the palaeosol fractions 0.2 - 2 µm decreases until 400 °C. Between 400 °C and 430 °C the intensity increases by about 1 % before it decreases again. At 620 °C the remanence disappears completely (Figure 5.5a, top). The magnetisation of the loess Li-L4 fraction 0.2...
- 2 µm decreases rather monotonously and vanishes also at 620 °C. The susceptibility of the three samples increases by 20 - 30 % between 300 °C and 600 °C, probably due to the formation of new remanence carriers.

The intermediate coercivity component residing in the fractions 5 - 20 µm and

**Figure 5.5a:** Thermal demagnetisation of IRM and susceptibility for the three different coercivity windows. The curves consist of 32 to 35 demagnetisation steps. **Top:** The grain size fractions of both palaeosols (Li-S1, Li-S5) which have magnetisations regarded to be of pedogenic origin, behave very similarly. The IRM of the same grain size fraction of loess Li-L4 differs to some extent but its susceptibility is very similar to that of the palaeosols. Maximum unblocking of IRM occurs at ~620 °C suggesting the presence of maghemite. **Middle:** Thermal demagnetisation of the two loess grain size fractions is slightly different. The remanence disappears above 665 °C and indicates the presence of hematite. A substantial amount of magnetic minerals is formed during heating (susceptibility increase). **Bottom:** Hematite is regarded to be the high coercivity remanence carrier. Its remanence vanishes at 685 °C in both samples. The increase in susceptibility indicates formation of new ferromagnetic minerals, but they do not acquire remanence. The susceptibility of By-L4 has not been measured.
20 - 50 µm of the two loess samples shows similar thermal behaviour (Figure 5.5a, middle panel). The remanence being unchanged until 100 °C disappears at 675 °C. The rapid loss up to 250 °C is stronger in the finer fraction. Another rapid loss occurs at 650 °C in both fractions. The susceptibility starts to increase by about 60 - 80 % between 400 °C and 500 °C, indicating the formation of new magnetic minerals.

The high coercivity components of the pristine loess bulk sample By-L4 and the well developed palaeosol bulk sample Li-S1 decay quasi-linearly up to 660 °C (Figure 5.5a, bottom). Above this temperature the magnetisation decreases very rapidly and disappears at 685 °C. The susceptibility increases by about 30 % between 300 °C and 600 °C.

The thermomagnetic curves of the loess samples Li-L4 and Li-L8 (fraction 5 - 20 µm) are identical (Figure 5.5b, left, upper panel). Paramagnetic behaviour predominates from –196 °C to room temperature. The inflection between 350 °C and 500 °C may indicate destruction of a thermally unstable phase, possibly maghemite. The Curie temperature is about 610 °C. A second heating cycle shows hardly any ferromagnetic contribution. The thermomagnetic behaviour of both loess fractions 20 - 50 µm (Figure 5.5b, right, upper panel) shows reduced paramagnetic influence below room temperature. The magnetisation decreases
steadily and without inflection between 350 °C and 500 °C until 610 °C. The considerable loss of magnetisation below this temperature indicates the presence of magnetite. The second heating differs from the first heating indicating destruction of maghemite. The Curie temperature curves of the palaeosol grain size fraction do not show paramagnetic behaviour. The magnetisation decreases quasi-linearly and disappears at Curie temperatures of 575 °C and 625 °C in palaeosol Li-S1, fraction 0.2 - 2 µm, respectively. (The second heating curve of Li-S1 is quite different and indicative for magnetite, which was formed during the first heating from previous maghemite contributions. This unusual reaction is due to the wax, which provides a reducing carbon dioxide/monoxide atmosphere.)

The susceptibility of the Li-S1 and Li-S7 grain size fractions (Table 5.1) generally decreases with increasing grain size except for the smallest fractions. Maximal values are observed in the fractions 0.2 - 2 µm of both palaeosols, minimal values in the fractions > 50 µm. The susceptibility of Li-S1 fractions is always higher than in the fractions of Li-S7. A direct relation between susceptibility and grain size is not recognised in the loesses. The most prominent susceptibilities are found in fraction 20 - 50 µm with slightly higher value in Li-L8 than in Li-L4. The susceptibility of the smallest fraction is also enhanced, whilst that of fraction 5 - 20 µm is constant. The values of all other fractions are enhanced in Li-L4.

The total iron content of the different grain size fractions varies between 1.7 and 8.6 % (by weight) (Table 5.1) and is in the smaller fractions generally higher than in the coarser fractions. A direct relationship between the iron content of the grain size fractions and their susceptibility does not exist. Regarding palaeosols, increased susceptibility values correlate with increased
iron contents only, if the two smallest fractions are subsumed. This is not the case in the loess where maximal susceptibility anti-correlates with lowest iron content (fraction 20 - 50 μm). The titanium content varies little from 0.16% to 0.65 % (by weight). The highest concentrations are observed in the intermediate grain size fractions (2 - 5 μm and 5 - 20 μm), low values in the fractions < 0.2 and > 50 μm. No correlation with susceptibility is recognised.

Figure 5.6: Four coercivity components have been chosen as endmembers for a linear mixing model for loess and palaeosol samples to model the magnetisation of a bulk sample. The measured endmember spectra (line with error band) have been modelled using a linear combination of three lognormal distribution functions (black line) – except component H, where one function was used – in order to get a best mathematical approximation to the observation. The spectra have been truncated (dashed line), because it is assumed that they consist of a single component only. In all cases the best fit, obtained using the Levenberg-Marquardt (Marquardt, 1963) minimisation algorithm of “Mathematica”, is within the error of the measured spectra.
5.5 Linear source mixing model

The IRM acquisition spectra of the grain size fractions (Figures 5.3, 5.4) demonstrate the coexistence of different coercivity populations as already suggested by the IRM/ARM demagnetisation spectra of the bulk samples (Figure 5.2). In order to quantify the magnetisation contributions, a linear source mixing model was developed using principal component analysis.

The four different coercivity populations have been characterised in those grain size fractions where they are most prominently developed. Component P has been obtained from the spectra of Li-S1 (0.2 - 2 µm fraction), component D1 from Li-L4 (5 - 20 µm fraction), component D2 from Li-L8 (20 - 50 µm fraction) and component H from Li-L4 (< 0.2 µm fraction). In order to obtain the most likely mathematical approximation of these components (endmembers), the spectra were modelled by a linear combination of a certain number of lognormal functions (Figure 5.6). Component H was modelled using only one lognormal function. The modelled spectra (black lines) are nearly congruent to the observed ones (white lines) and are always within the errorband (grey bands) of the observed LCS’s. Although modelled and observed spectra of component H are not congruent, the model curve still runs within the errorband and is considered to be reliable and representative.

The IRM acquisition curves of the grain size fractions have been stacked (taking into account the mass contribution of the different grain size fractions in order to model the IRM acquisition curves, which represent the bulk samples (Figure 5.7). All characteristics of the modelled endmembers (mean coercivity, dispersion, general shape) have been determined by the observed spectra of the grain size fractionated samples and have been kept constant at this step. The individual contribution to the total IRM was obtained by integration (Table 5.3). Small deviations are observed at the low coercivity part of both palaeosols, at the maximal LCS amplitude of palaeosol Li-S1 and at the part between 300 and 600 mT of the loesses.

The remanence of component P predominates clearly the total IRM of palaeosol Li-S1 with 66 %, whereas P contributes about 44 % only in palaeosol Li-S7. Component D1 predominates with 52 % loess Li-L4 and component D2 with 58 % loess Li-L8. Components H and D1 can be considered as constant in all four samples. They amount from 0.35 to 0.49 mAm²/kg and from 1.28 to 2.36 mAm²/kg, respectively. Components P and D2 have indeed a rather variable
contribution from 0.10 to 9.09 mAm$^2$/kg and from 0.75 to 3.05 mAm$^2$/kg, respectively.

**Figure 5.7:** Modelling the bulk IRM spectra using a linear combination of the four endmembers (black line) P, D$_1$, D$_2$ and H discussed in Figure 5.6. The white curve with the grey errorband represents the bulk spectrum of the sample. The loess samples Li-L4 and Li-L8 can be well represented by the model and the best fitting linear combination is still within the error band of the measured spectra. The best fit deviates slightly between 300 and 600 mT in the loesses but also in the low coercivity part of the palaeosols, where possibly a fifth component with very low coercivity could be placed. The area below each component represents its individual contribution to the bulk magnetisation of the sample and was obtained by integration (compare Table 5.3).
Interpretation

ARM and IRM respond to different magnetic domain stages. ARM focuses specifically on single domain (SD) grains (Dunlop and Özdemir 1997; Egli and Lowrie, 2002), whereas IRM magnetises all possible remanence carriers. This explains the generally smeared appearance of the maxima in the IRM spectra of the low intensity loesses. The observed LCS maxima of ARM and IRM demagnetisation curves are comparable (cf. Figure 5.2) as both magnetisations have been demagnetised by AC fields.

Component P is best documented in all experiments. It peaks in the ARM demagnetisation curves at 22 mT, in the IRM demagnetisation spectra at 20 mT and in the IRM acquisition spectra at 31 mT (Figure 5.8). The value of 31 mT agrees with those of Eyre (1996) using IRM acquisition, too. He reports a mean

### Table 5.3: Absolute and relative contribution of the four main remanence carriers to the bulk magnetisation of the analysed loesses and paleosols. Component P increases with increasing pedogenesis, whereas component D1 and H are rather constant. Component D2 has also variable contributions. The errors are estimated by fitting the upper and lower error boundary (cf. Figure 5.7) of the bulk spectra.

<table>
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<tr>
<th>Horizon</th>
<th>Li-L4</th>
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</thead>
<tbody>
<tr>
<td>Contribution</td>
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<td>relative</td>
</tr>
<tr>
<td>mAm²/kg</td>
<td>%</td>
<td>mAm²/kg</td>
</tr>
<tr>
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<td>D2 (79 mT)</td>
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<td>D1 (113 mT)</td>
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<td>H (610 mT)</td>
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<td>IRM at 4.5 T</td>
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coercivity of (34.8 ± 8.4) mT (95% c.l.) for 41 analysed loess/palaeosol samples. These values are close to Maher’s (1988) results for synthetic single domain magnetite. She found values of 21 – 29 mT using IRM acquisition.

The thermal demagnetisation of IRM and the Curie curves suggests the presence of slightly oxidised magnetite forming component P. Magnetite with a grain size of about 50 - 60 nm is the main ARM carrier (Dunlop and Özdemir 1997; Egli and Lowrie, 2002). Chemically grown grains of biogenic or abiotic origin, which may have formed during pedogenesis, have such small grain sizes. A biogenic intracellular origin is excluded because the ARM$_{DC=0.1mT}$/SIRM ratio of palaeosol Li-S1 which is predominated by the pedogenic component P is about 0.06. This is much lower than the values reported by Moskowitz et al. (1993) for intracellular magnetite (intact magnetosomes) (ARM$_{DC=0.1mT}$/SIRM = 0.15 – 0.25). Magnetotactic bacteria need specific oxygenation conditions to grow magnetosomes, which probably did not prevail in the “dry” palaeosols on the CLP. Favourable conditions would rather be met in stagnant and/or poorly drained soils with varying water level (Fassbinder et al., 1990) and oxygen-poor water. Redoximorphic features, such as Fe-Mn films, concretions and mottling have not been observed in palaeosol S1 and only rarely in palaeosol S7 (Ding et al., 1999). Since the spectra of Li-S1 and Li-S7 are almost equal (see Figure 5.3, grain size fraction 0.2 - 2 µm), the formation of magnetosomes has been excluded. Maher and Thompson (1995) show evidence for magnetic particles in palaeosols, which are similar to magnetosomes and suggest that they play a minor role in the magnetic enhancement of palaeosols. This does not mean that biologically mediated iron oxides do not exist generally in loess/palaeosols. Chukhrov et al. (1976) found that Fe$^{3+}$-oxide formed by bacterial activity in soils is usually ferrihydrite. According to Schwertmann (1988b), an inorganic formation of magnetite from mixed Fe$^{2+}$/Fe$^{3+}$ solutions at room temperature under neutral or alkaline pH is possible via (proto-)ferrihydrite or other trivalent Fe oxides such as lepidocrocite (Schwertmann and Taylor, 1973).

We suggest production of the pedogenic component in the following way: The weathering of detrital parent material provides a solution of bi- and trivalent iron from which a nonmagnetic protocompound of trivalent iron is built up. A catalysator in form of a magnetite nucleation germ is needed to force magnetite crystallisation and growth under these conditions. Eventually chemical changes of the soil solution may lead to the development of a diffuse fringe of protocompounds such as ferrihydrites surrounding the growing magnetite grain.
Pure pedogenic magnetite, however, is difficult to verify in soils (Schwertmann, 1988b), because it undergoes easily partial oxidation. Hence we assign pedogenic maghemite of low oxidation state as carrier of component P.

High unblocking temperatures near 685 °C during thermal demagnetisation of IRM (Figure 5.5a, bottom), high coercivity (Figure 5.8) and low magnetisation indicate the presence of hematite as a carrier of component H. The grain size is

Figure 5.8: A possible mineralogical interpretation of the four main coercivity populations (endmembers) coexisting in loess/palaeosol samples. The high coercivity component H is due to the presence of hematite. Its coercivity of 610 mT suggests a small grain size with an upper limit of 2 μm (see text). Because the contribution varies very little between samples (Table 5.3), its origin is most probably detrital. Component D1 is also considered to be of detrital origin because it varies within fairly narrow limits. The coercivity at 113 mT would assign maghemite of high oxidation stage. Component D2 has a rather variable contribution with a significantly lower coercivity than D2 and is interpreted as altered detrital magnetite. The content of component P with a coercivity of 31 mT varies largely. It is of pedogenic origin and represented by magnetite or maghemite in the single domain state.

(Dearing et al., 1996).
below 5 µm according to coercivity data of Dankers (1981). Dunlop’s (1981) values suggest a grain size of 0.2 - 1.0 µm within the single domain range of equidimensional hematite. The remanence contribution of component H is quite constant in all samples including pristine loess Li-L4. Hence, we favour a detrital origin and suggest that component H is due to fine-grained single domain, detrital hematite with an upper grain size limit of about 2 µm. The reddish colour of the palaeosols is caused by in-situ grown hematite grains. Their size ranges between 0.01 and 0.1 µm (Schwertmann, 1988a, Scheinost and Schwertmann, 1999) and is close to superparamagnetic grain size. Trivalent iron may originate from hydrolithic and oxidative decomposition of lithogenic bivalent iron silicates, such as fayalite (Schwertmann, 1988b). Depending on pedoenvironmental conditions such as temperature, rainfall and organic matter, the trivalent iron can either form goethite by crystallisation or ferrihydrite via de-protonisation and hematite via further dehydration and structural rearrangement (Schwertmann, 1988b). The red palaeosol colouration due to pedogenic hematite (Scheinost and Schwertmann, 1999; Chen et al., 2002) does not contribute to IRM and ARM. This is in agreement with the finding of Vandenberghe et al. (1998) using Mössbauer spectra of loess/palaeosols samples from the central CLP.

Component D1 is observed in the IRM demagnetisation spectra of palaeosols and loesses, but not in the ARM spectra. The analysis of the IRM acquisition spectra indicates coercivities around 113 mT (Figure 5.8). This excludes single domain magnetite as a potential remanence carrier of that component. In addition, thermal demagnetisation of IRM and Curie temperatures around 610 °C (Figure 5.5b) point to the presence of maghemite in loess grain size fraction 5 - 20 µm which is prominent for component D1. We interpret component D1 as detrital maghemite because it dominates the bulk magnetisation of the pristine loess Li-L4 with 52 % and contributes rather constantly in the different samples. This is in agreement with earlier findings (Heller and Liu, 1984; Evans and Heller, 1994) which suggested the presence of detrital maghemite, being resistant against weathering.

A similar maghemite phase with Curie temperatures of 610 °C is found in the loess grain size fraction 20 - 50 µm. Magnetite is also present, because the thermomagnetic curve is not inflected between 300 and 500 °C. The component D2 with LCS maximum at 79 mT (Figure 5.8) characterises the grain size fraction 20 - 50 µm in the weathered Li-L8. D2 occurs in all samples but with
highly variable amounts. Therefore D₂ is interpreted as an altered magnetite or as a detrital magnetite with a weathered crust of maghemite.

The thermomagnetic curves, the IRM acquisition spectra and the presence in the ARM demagnetisation spectra show significant differences between component D₂ and D₁ although their LCS maxima are close to each other. Since model bulk spectra, which have been calculated using only one of the components, are dissimilar, the measurement curves D₁ and D₂ are considered to reflect different components.

The linear mixing models of the palaeosols deviate slightly from the measured curves in the very low coercivity part (Figure 5.7, palaeosols). A fifth component (L) may be present with a peak between 13 mT (Figure 5.2, lower part) and around 21 mT (Figure 5.3). This component could correspond to Eyre's component 1 peaking at (10.7 ± 3.8) mT. Component L would be attributed to magnetite grains (Curie temperature of 575 °C, see Figure 5.5b, lower left) below the stable single domain size, but large enough to carry stable remanences. It is probably of pedogenic origin because it is much more pronounced in palaeosols than in loesses (Figure 5.4).

Finally, the remanence contributions of the individual components might be compared cautiously with the total iron content of the grain size fractions. The pedogenic component P, which is strongly developed in the palaeosol grain size fraction 0.2 - 2 µm, is related to high iron content (Figure 5.9). The iron content is also high in this grain size fraction of the loesses, but P has not been developed yet. Hence, iron is exchanged during pedogenesis between detrital non-magnetic ferruginous minerals and new pedogenic iron oxides. The cross iron content remains unchanged.

Clay mineral assemblages of Chinese loess sediments are dominated by iron-bearing clays such as illite and chlorite (Liu, 1985). During pedogenesis the clays may release iron and nanometer-sized ferromagnetic minerals crystallise between clay minerals. The susceptibility of the clay sized fraction increases. This has been confirmed recently by Ji et al. (2002) who observed nanometric magnetite, hematite and goethite crystal coating on clay minerals.

Component D₁ correlates with the high titanium content of the 5 - 20 µm fraction (Figure 5.9). The chemical immobile titanium is of detrital origin and this supports the arguments for a detrital origin of D₁ itself. The titanium may reside in rutile or augite rather than in titanomaghemite for which no experimental evidence was found.
Different magnetic methods in combination with grain size fractionation identify the coexistence of four coercivity populations in loess/palaeosol samples from Lingtai (central Chinese Loess Plateau). The two-component model by Evans and Heller (1994) has been extended to four components similar to the model of Eyre (1996). Using the argument of the linear additivity of remanences a mixing model based on the observed coercivity spectra of certain grain size fractions was used to model bulk IRM acquisition coercivity spectra of loess/palaeosol samples. The following conclusions can be drawn:

- Each loess/palaeosol sample contains at least four different coercivity populations (P, D₁, D₂, H), which are able to carry stable remanence. The minerals magnetite, maghemite of low and high oxidation state and hematite have been identified as remanence carriers.
- The contribution of the different coercivity components varies with lithology and has been quantified. The pedogenic low coercivity component P predominates the palaeosols with variable contributions of 44 to 66 %, whereas the detrital intermediate coercivity component D₁ is rather constant in all samples. It is most important in pristine loess with 52 %. The intermediate coercivity component D₂ contributes variably, but gives a main contribution in weathered loesses (58 %). A constant detrital hematite component is also present in all samples with contributions up to 10 %.
- The detrital remanence carriers (H and D₁) have not been destroyed during weathering. Their contribution is rather constant in loesses and palaeosols. Hence, the iron for the new pedogenic (remanence carrying) component in palaeosols and weathered loesses must derive from clays and other ferruginous silicates.
- The overall iron content of the clay-sized fraction does not change during pedogenesis but the ferromagnetic mineral content increases strongly. The pedogenic iron oxides grow within this grain size fraction on translocated clays. The iron brought into solution by pedogenesis apparently re-precipitates after short migration.
- The pedogenic component P increases with the degree of pedogenesis, similarly as the bulk susceptibility (Table 5.1 and 5.3). Hence component P indicates climate variations, both in time and space, which cause variable in situ alteration of the loess material mainly by differing palaeoprecipitation (cf. Heller et al., 1993; Maher and Thompson, 1995). In contrast, variations of component D₁ might reflect variable aeolian sedimentation conditions caused by different wind regimes.
Model

A lock-in model for the complex Matuyama-Brunhes boundary record of the loess/palaeosol sequence at Lingtai (Central Chinese Loess Plateau)

Summary

In most marine sedimentary records, the Matuyama-Brunhes boundary (MBB) has been found in interglacial oxygen isotope stage 19. In the magnetostratigraphic records of most Chinese loess/palaeosol profiles the MBB is located in loess layer L8, which was deposited during a glacial period. The MBB at Lingtai (Central Chinese Loess Plateau) also occurs in L8 and is characterised by multiple polarity flips. The natural remanent magnetisation is mainly carried by two coexisting components. The higher coercivity (harder) component dominates in loess sections and is thought to be of detrital origin. The lower coercivity (softer) component prevails in palaeosols and was most probably formed in-situ by (bio-)chemical processes. A lock-in model for the Lingtai MBB record has been developed by extending the lithologically controlled PDRM model of Bleil and von Dobeneck (1999). It assumes two lock-in zones. The NRM of the magnetically harder component is physically locked by consolidation shortly after loess deposition, whereas the softer component is formed at greater depth by pedogenesis and acquires a chemical remanent magnetisation of younger age. At polarity boundaries, grains carrying reversed and normal directions may therefore occur together within a single horizon. The model uses ARM coercivity spectra to estimate the relative contributions of the two components. It is able to explain the observed rapid multiple polarity flips and low magnetisation intensities as well as the stratigraphic shift of the Lingtai MBB with respect to the marine records.

Keywords: Detrital remanent magnetisation, chemical remanent magnetisation, Matuyama-Brunhes boundary, loess, palaeosol, China.
6.1 Introduction

The Matuyama-Brunhes boundary (MBB) has been observed in many sediment cores from different oceans. Tauxe et al. (1996) compiled 19 such records with sedimentation rates of up to 8 cm/kyr and confirmed that the MBB occurs in most cores in sedimentary layers representing the interglacial oxygen isotope stage 19. Astronomical calibration of the marine oxygen isotope record places the MBB at 778.8 ± 2.5 ka (Tauxe et al., 1996). This date agrees well with radiometric ages derived from volcanic sequences, which give 778.2 ± 3.5 ka (Tauxe et al., 1996) or 780 ± 10 ka (Spell and McDougall, 1992) as the best estimate. Tauxe et al. (1996) also reported that the natural remanent magnetisation (NRM) in marine sediments is fixed within a few centimetres of the water – sediment interface. Among others, deMenocal et al. (1990) and Bleil and von Dobeneck (1999), report downward shifts of magnetic polarity boundaries in marine sediments, with values ranging of 7 to 17 cm.

Surface layers of marine sediments are commonly unconsolidated due to high porosity and frequent bioturbation in the upper 5 - 15 cm. An acquired depositional remanent magnetisation (DRM) cannot persist under these conditions (Guinasso and Schink, 1975). According to the widely accepted concept of a post-depositional remanent magnetisation (PDRM), the magnetisation is gradually locked below this unstable layer after a certain amount of overburden has accumulated on the palaeosurface (Irving and Major, 1964; Kent, 1973; Otofuji and Sasajima, 1981). Because the post-depositional fate of each magnetic particle is of a statistical nature, any magnetic assemblage will exhibit a distribution of different individual lock-in depths - even under the most idealistic assumption of uniform magnetic and non-magnetic matrix particle sizes. The macroscopic process is therefore appropriately described by a ‘lock-in zone’ (Løvlie, 1976; Niitsuma, 1977) rather than by a sharp ‘lock-in front’. Consequently, lock-in not only delays remanence acquisition with regard to sediment age, but it also acts as a smoothing filter in the depth and time domains. Effects of delayed and gradual recording upon the palaeomagnetic fidelity of homogeneous sediments have been mathematically investigated by Denham and Chave (1982) and Hyodo (1984).

Bleil and von Dobeneck (1999) defined the initial lock-in depth \(d_0\) as the depth above which no lasting magnetisation is acquired. Under simple assumptions, a magnetic polarity boundary appears in the record where 50% of the magnetic
moments have been blocked parallel to the old polarity, and 50% of the moments are blocked parallel to the new field polarity. The corresponding depth is called the **median lock-in depth** \(d_{1/2} \). The depth where all available magnetic particles have been fixed is the **total or terminal lock-in depth** \(d_t \).

If lock-in is regarded as a steady-state process, brief polarity intervals will be recorded only if they last longer than the median lock-in depth divided by the sedimentation rate. Therefore, short-lived geomagnetic events may not impart a clear polarity imprint on a PDRM record.

Terrestrial loess sediments also record geomagnetic field behaviour. Loess sequences consist of loess/palaeosol alternations where loess layers are relatively fresh aeolian deposits formed during colder climate periods, whereas palaeosols develop on a loess layer by pedogenic processes during warmer and wetter conditions. Heller and Liu (1982, 1984) observed the MBB at Luochuan (central Chinese Loess Plateau, CLP) in the 8th palaeosol (palaeosol S8) from the top of the sequence. Many other authors observed this boundary in the more recently deposited loess layer above S8 (called L8 according to the nomenclature of Liu and Chang, 1964). Sometimes the MBB was not noticed at all in this entire stratigraphic interval (Hus and Han, 1992). Hus and Han (1992) pointed out that different PDRM lock-in depths could explain the different stratigraphic positions of the MBB.

Variability of lock-in depths may be due to different post-depositional processes. The fixing of magnetic grains in loess depends on the stabilisation of the sediment microstructure. After the settling of the dust particles, "loessification" (transformation of dust into loess coupled with secondary calcification) establishes this microstructure. This structural change takes place in a depth zone between ~0.4 and < 1 m under specific loading and wetting conditions (Assallay et al., 1998). In addition, the presence of oxidised titanomagnetites (Heller and Liu, 1984), the alteration of unstable minerals (biotite, augite) in strongly weathered loesses (Liu, 1985) and the reductive dissolution of iron which can be reprecipitated in oxidising or carbonate-rich environments (Perel’man, 1977) would also argue for the formation of a chemical or crystallisation remanent magnetisation (CRM). There is now overwhelming evidence that the MBB on the Chinese Loess Plateau is recorded mostly in loess layer L8 (Liu et al., 1988; Cao et al., 1988; Rolph et al., 1989; Rutter et al., 1990; Zheng et al., 1992; Li et al., 1997; Ding et al., 1998; Zhu et al., 1998; Spassov et al., 2001).
Based on a detailed stratigraphic comparison between neighbouring loess/palaeosol sections of the central CLP, Heller et al. (1987) suggested local relative remanence lock-in delays of about 20 kyr for the MBB and proposed a correlation of palaeosol S7 with marine oxygen isotope stage 19. Zhou and Shackleton (1999) compared the MBB records of deep-sea cores and other (lacustrine) continental sediments and concluded that the MBB should be observed within oxygen isotope stage 19. They considered the MBB position in the marine sediments as a reliable time marker and argued that the MBB in the loess sediments at Luochuan, the classical section on the CLP, was displaced downward by 1.7 to 2.5 m and that oxygen isotope stage 19 should be correlated with palaeosol S7. They used the 700 – 800 ka old Australasian micro tektite layer, which was observed in deep-sea cores about 12 kyr before the MBB (Schneider et al., 1992; Kent and Schneider, 1995), to fix their correlation in an absolute time frame. Some microtektites have also been identified in the upper part of L8, but above the recorded MBB (Li et al., 1993).

The Chinese loess/palaeosol timescale developed by Heslop et al. (2000) is based on the correlation of astronomically tuned monsoon records with the oxygen isotope record of ODP site 677 (Shackleton et al., 1990). Heslop et al. (2000) concluded that the polarity boundaries are displaced and shifted downward in the loess stratigraphy. They arrived at shift estimates of 1.90 m at Baoji (central CLP) and 1.59 m at Luochuan, which would correspond to time delays of 26 kyr and 23 kyr, respectively, and they also suggested a correlation of palaeosol S7 with marine oxygen isotope stage 19 (see also Evans and Heller, 2001).

The present study is intended to shed some light on the processes, which cause the apparent time lag of the MBB in the Chinese loess sediments with respect to the marine record. The loess section at Lingtai (central CLP) has been selected for detailed rock- and palaeomagnetic measurements. Post-depositional and chemical processes of remanent magnetisation acquisition will be considered in the development of a lock-in model, which extends the PDRM model proposed by Bleil and von Dobeneck (1999).
6.2 Magnetic polarity boundaries and rock magnetic properties of the Lingtai section

The Lingtai section (34.98° N, 107.56°E) is located in the central part of the CLP. The upper 175 m of the 305-m-thick section consists of alternating loess/palaeosol and the lower part is formed by Pliocene red clay with a basal age of about 7.05 Ma (Ding et al., 1998). Some 32 palaeosols have been identified in the Lingtai sequence. They can be correlated with other sections, for instance with Baoji (Rutter et al., 1991), and provide a complete stratigraphic sequence. The Lingtai section was sampled continuously from 0 to 268.12 m, resulting in about 13, 400 oriented cubic samples with an edge of 2 cm.

Here we discuss results from samples spanning the MBB. A first sample set (set A) was collected to provide continuous data for a 3 m stratigraphic interval from S8 through the slightly weathered L8 up to S7. All samples from set A were stepwise thermally demagnetised. Another independent sample set (set Y, collected about 1 m away from set A) was treated using alternating fields (AF) demagnetisation. A secondary present-day overprint was removed at temperatures between 250°C and 300°C or at peak fields of 15 mT or 20 mT in both loess and palaeosol samples. After demagnetisation treatment the MBB was found between 61.4 and 62.0 m in L8 (Figure 6.1), implying an average sedimentation rate of 7.9 cm/kyr for the Brunhes part of the section.

The stable, characteristic remanent magnetisation (ChRM) records seven full polarity changes at the MBB throughout a transitional zone of about 0.4 m thickness. They occur at slightly different depths in the two data sets (Figure 6.1). Two intermediate VGP positions are observed in the Y data (AF) within 0.5 m below the lowermost full reversal. They are not present in set A (thermal). The partly dissimilar ChRM directions seem to reflect different demagnetisation response of the coercivity and the blocking temperature distributions of the magnetic mineral fractions present.

In order to assess the reliability of the data, the maximum angular deviation (MAD) of the regression line of the ChRM was calculated, using the same number of demagnetisation steps (6) for all samples (Spassov et al., 2001). Both demagnetisation methods yield approximately the same precision above and below the transition zone with MADs generally < 10° whereas the ChRM directions are less well defined within this zone, with MADs sometimes exceeding 30°. This may be due to increasingly mixed ChRM polarity within
these samples. The two intermediate VGPs also have high MAD values (Spassov et al., 2001).

Specific low-field susceptibility $\chi$, coercivity $H_c$, coercivity of remanence $H_{cr}$, saturation magnetisation $M_s$ and saturation remanence $M_{rs}$ vary smoothly across this stratigraphic section and do not show any abrupt changes near, or within, the transition interval (Figure 6.1). The coercivity values slightly increase from S8 to L8 and suggest that two coercivity populations, coexist in the loess/palaeosol samples (cf. Evans and Heller, 1994). Higher concentrations of a
rather fine-grained mineral fraction (magnetite) are indicated in palaeosol S8 by higher frequency dependence of susceptibility (F-factors) and saturation magnetisations together with lower coercivities. We argue that the multiple polarity changes, which do not occur completely simultaneously in both data sets, were caused by lock-in effects rather than by geomagnetic field behaviour (see Spassov et al., 2001). They are not correlated with any distinct changes of the rock magnetic signature or lithology. The syn- or post-sedimentary formation of detrital and pedogenic minerals, respectively, may have had an important influence on the lock-in process of the ChRM.

6.3 IRM analysis

Detailed analysis of acquisition and demagnetisation of isothermal remanent magnetisation (IRM) data can provide critical information about coercivity distributions and related mineral phases (Robertson and France, 1994; Kruiver et al., 2001; Egli, 2003). Five samples were selected to investigate the influence of weathering and pedogenesis on the magnetic mineralogy of the loess samples using the technique of Egli (2003). The first sample (BY055) is from the L4 (depth = 69 m) of the high sedimentation rate Baicaoyuan section in the western CLP. It has a low susceptibility of 38x10⁻⁸ m³/kg and is regarded as being essentially unaltered and unaffected by weathering (see Evans and Heller, 1994). The second sample (A1515, depth = 30.3 m) originates from L4 of the Lingtai section. Although this sample is from the central CLP, where sedimentation rates were lower and weathering and pedogenic alteration is generally stronger, it has an even lower susceptibility of only 21x10⁻⁸ m³/kg. The third sample (Y0030) is from the upper part of L8 (depth = 60.9 m) at Lingtai and has a susceptibility of 26x10⁻⁸ m³/kg. The most mature stage of pedogenesis within the interval of interest is expected to be seen at Lingtai in sample A2980 from the weakly developed palaeosol S7 (depth = 59.6 m). Its susceptibility is 77x10⁻⁸ m³/kg. Sample A0494 originates from one of the most strongly developed palaeosols at Lingtai – palaeosol S1 (depth 9.88 m). It has a susceptibility of 215x10⁻⁸ m³/kg.

Preliminary IRM acquisition curves up to 4600 mT saturate at about 3000 mT for loess samples and at 1500 mT for palaeosols. Since the experiments described below involved IRM’s acquired in fields up to 300 mT we will be
concerned only with the low coercivity mineral content. This seems to be acceptable as the increase of the IRM acquisition curve above 300 mT is small in both lithologies. In palaeosol and loess samples 93% and 85% of the saturation remanence are reached at 300 mT, respectively.

All samples were first demagnetised using an AF of 300 mT along three orthogonal axes. The samples were then magnetised with a DC field of 300 mT along one axis. Tests on samples with different amounts of viscous remanence carriers show that after a waiting time of about 3 minutes almost no decay of the IRM was observed within the time required for the measurement to be completed. After this delay, stepwise AF demagnetisation (logarithmic steps up to 300 mT) of the IRM was performed along the magnetised axis was started.

**Figure 6.2:** (a/b) AF demagnetisation of IRM given at 300 mT for different loess/palaeosol samples. (c/d) Gradients of the demagnetisation curves. The higher coercivity component (maximum 40 - 58 mT) is present in all samples (even in the strongly developed palaeosol S1) and represents the detrital population D, assumed to be uniform across the Chinese Loess Plateau (Evans and Heller, 1994). With increasing humidity the loess is affected by weathering and two new lower coercivity populations P of (bio)chemical grains is formed. The contribution of each component is given in Table 6.1. The dashed line in (d) represents the continuation of the spectrum of palaeosol S7.
and the remaining IRM was measured using a 2G Enterprises cryogenic magnetometer with an in-line AF demagnetisation coil. The differential IRM demagnetisation curve was calculated using the following steps. First, the original curve was scaled resulting in a linear demagnetisation curve. Next, the scaled demagnetisation curve was fitted with a hyperbolic tangent function.

Then the residuals between the fit and the data were low-pass filtered (using the same Butterworth filter parameters (order 8) for all demagnetisation curves) in order to eliminate experimental noise. After backward transformation, a noise-free IRM demagnetisation curve was obtained. This curve was then scaled using a logarithmic field scale (base 10), thus the field axis becomes unitless. The resulting derivative, also called the logarithmic coercivity spectrum (LCS), therefore has the same units as the magnetisation (for a complete description of the method see Egli, 2003).

The initial IRM demagnetisation curves show progressively higher intensities with increasing degree of weathering (Figure 6.2a, b). All IRM gradient curves exhibit a coercivity component that peaks between 40 and 58 mT, which is attributed to minerals of detrital origin (Figure 6.2c, d). A lower coercivity maximum develops progressively from L4 at Baicaoyuan (BY055) to L8 at Lingtai (Y0030) into the palaeosols S7 and S1. The low coercivity peak is located between 16 and 25 mT. This peak appears to be caused by pedogenic processes as demonstrated by its dominance in the spectra from palaeosols S7 and S1. The higher coercivity component of S7 (dashed line in Figure 6.2d) has essentially the same amplitude as that of the parent material in L8.

The different magnetisation contributions were quantified following the unmixing method proposed by Egli (2003). Magnetisations of mixed components add linearly (Stacey, 1963; Kneller and Luborsky, 1963; Roberts et al., 1963). The pedogenic component P increases gradually from the relatively unaltered loess BY055 on the western CLP representing cold/arid climate conditions to the altered loess Y0030 on the central CLP where more humid climate conditions prevailed. Maximum pedogenesis within the interval of interest is reached in the palaeosol sample A2980.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Horizon</th>
<th>Location</th>
<th>Contribution of</th>
<th>Initial IRM</th>
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<tbody>
<tr>
<td></td>
<td></td>
<td>on CLP</td>
<td>P</td>
<td>D</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>mAm²/kg</td>
<td>mAm²/kg</td>
</tr>
<tr>
<td></td>
<td></td>
<td>western</td>
<td>0.43 (11.5%)</td>
<td>3.27 (88.5%)</td>
</tr>
<tr>
<td>BY055</td>
<td>L4</td>
<td>central</td>
<td>0.40 (9.8%)</td>
<td>3.67 (90.2%)</td>
</tr>
<tr>
<td>A1515</td>
<td>L4</td>
<td>central</td>
<td>0.52 (11.2%)</td>
<td>4.12 (88.8%)</td>
</tr>
<tr>
<td>Y0030</td>
<td>L8</td>
<td>central</td>
<td>2.94 (47.1%)</td>
<td>3.30 (52.9%)</td>
</tr>
<tr>
<td>A2980</td>
<td>S7</td>
<td>central</td>
<td>6.24</td>
<td>6.25</td>
</tr>
</tbody>
</table>

**Table 6.1:** The pedogenic component P increases gradually from the relatively unaltered loess BY055 on the western CLP representing cold/arid climate conditions to the altered loess Y0030 on the central CLP where more humid climate conditions prevailed. Maximum pedogenesis within the interval of interest is reached in the palaeosol sample A2980.
Therefore the model attributes the coercivity spectra to a number of distinct coercivity populations. Our analysis indicates the presence of one detrital (\(D\)) and one pedogenic (\(P\)) mineral population. The spectrum for sample BY055 (supposed to represent the pristine loess end-member, see Figure 6.2) was subtracted from the spectrum of palaeosol S1 (supposed to represent the pedogenic end-member, see Figure 6.2) in order to obtain the pedogenic component \(P\) in the sense: \(P = S1 - n \cdot BY055\) (\(n = 1.5\). With this factor, the detrital component disappears completely in the spectrum of S1). After fitting \(P\) with a log-normal function, it was subtracted again from BY055 to obtain the pure detrital component: \(D = BY055 - k \cdot P\) (\(k = 0.04\). With this factor the pedogenic component disappears in the spectrum of BY055). \(D\) was then fitted with log-normal functions. This algorithm was iteratively applied and a stable solution was reached after the second iteration step. Each pilot sample could be modelled using a linear combination of the two components. The area under each gradient curve represents its contribution to the total IRM of the sample. The percentage of the contribution was then obtained by integration of each individual component.

Whereas the detrital component is roughly constant, component \(P\) steadily increases from the almost unweathered loess BY055 and peaks in palaeosol S1 where pedogenic minerals show highest concentrations (Table 6.1). Thus, the contribution of population \(P\) varies depending on the stage of pedogenesis. It can also be concluded that the detrital component \(D\) is not much influenced by pedogenesis because its contribution in L8 and S7 is nearly equal. As fields up to 300 mT were applied, high coercivity minerals such as hematite or goethite contribute very little to the IRM gradient. The detrital population \(D\) (coercivity maximum between 40 and 58 mT) is interpreted as maghemite or titanomagnetite (cf. Eyre, 1996), whereas the population \(P\) (coercivity between 16 and 25 mT) probably consists of magnetite of chemical or biochemical origin. Referring to magnetite coercivities measured on IRM acquisition curves (Maher, 1988), the coercivity values for \(P\) are close to the magnetite single-domain/superparamagnetic boundary and single domain state, respectively. These attributes have to be considered with caution since AF demagnetisation of IRM and DC acquisition of IRM are physically different processes and hence result in different coercivity values (Cisowski, 1981; Fabian and von Dobeneck, 1997; Dunlop and Özdemir, 1997).
The IRM results confirm our initial assumptions about the degree of weathering of the different samples. The least altered loess sample (BY055) exhibits essentially no low coercivity peak $P$, but $P$ is evident in loess Y0030 due to weathering, i.e. increasing pedogenesis. The result of our rock magnetic investigation is in agreement with the results of Sartori (2000) concerning general grain size populations, which indicate preferential formation of small grains (clay size) during pedogenesis. Because the pedogenic magnetic component was formed by chemical diagenesis, it is likely to carry a CRM, which may have been acquired under conditions different to those of the detrital component.

6.4 Lock-in models
6.4.1 Detrital lock-in

The principles of delayed remanence acquisition have already been discussed, experimentally tested and modelled by Løvlie (1974, 1976), Guinasso and Schink (1975), Denham and Chave (1982), Hyodo (1984), Mazaud (1996) and Meynadier and Valet (1996). Recently, a delayed remanence acquisition model was re-designed and applied to Quarternary deep-sea sediments by Bleil and von Dobeneck (1999). To summarise briefly, a layer, which is now at depth $z$, was progressively buried under a sediment cover of thickness $\zeta$ with $\zeta(t) = 0 \ldots z$. During burial the magnetisations of oriented magnetic particles are gradually locked-in. The fraction of the locked magnetic moment can be described by a linear (or, alternatively, curvilinear) lock-in function $\lambda$ (Figure 6.3), which depends on burial depth and lithology. At the terminal lock-in depth $d_l$, 100% of the magnetic moments are locked. This lock-in depth changes with lithology, which is represented by a single parameter ($c$) or parameter set representing physical characteristics such as sedimentation rate, porosity, bioturbation, etc., which influence the lock-in properties of the sediment. Assuming lock-in to set in at the sediment surface, we have:

$$\lambda_{PDRM} = \lambda(\zeta, c(z)) = \begin{cases} \frac{\zeta}{d_l(c(z))} & \text{for} \quad \zeta < d_l(c(z)) \\ 1 & \text{for} \quad \zeta \geq d_l(c(z)) \end{cases} \quad (6.1)$$
When modelling the lock-in process, burial depth $\zeta$ is incremented in discrete steps $i=1...n$. The incremental magnetisation from $\zeta_{i-1}$ to $\zeta_i$ acquired at the present $z$ is proportional to the increment of the linear lock-in function $\Delta\lambda=\lambda[\zeta_i,c(z)]-\lambda[\zeta_{i-1},c(z)]$ called the lock-in rate. $H(z)$ is the intensity of the local magnetic field at the time when the layer at $z$ was deposited. Only two antiparallel field directions (normal and reversed) are assumed and introduced as positive and negative values of $H(z)$, respectively. The average magnetic field over the interval is $\frac{1}{2}[H(z-\zeta_i) + H(z-\zeta_{i-1})]$. The post-depositional remanent magnetisation $M$ is obtained by multiplying both terms and by summing up the partial remanences (arrows in Figure 6.4) of all $n$ steps to $\zeta_n = z$:

$$M(z) = \sum_{i=1}^{n} \left\{ \lambda_{PDRM}[\zeta_i,c(z)] - \lambda_{PDRM}[\zeta_{i-1},c(z)] \right\} \cdot \frac{H(z-\zeta_i) + H(z-\zeta_{i-1})}{2} \quad (6.2)$$

This equation can be written in analytical form:

Figure 6.3: Linear lock-in function $\lambda_{PDRM}(\zeta)$ for the PDRM model after Bleil and von Dobeneck (1999). The percentage of locked magnetisation is expressed as a function of burial depth $\zeta$. If the lock-in depth $d_i$ is reached, 100% of the magnetisation is blocked.
A polarity change will be recognised when 50% of the magnetisation is locked in the new field direction. Consequently, the polarity boundary is shifted downward by half the lock-in depth (Figure 6.4). From depth $z - \zeta_n = 0$ down to depth $z - \zeta_2 \leq d_1$, the polarity of the magnetic field is normal but not all the grains are fixed. Therefore, the total remanence signal does not reach its maximal value of 1. At depth $z - \zeta_1$, all magnetic grains are fixed, but as the polarity changes the contributions almost cancel each other. The reversal is observed at the depth where the total remanence is zero (slightly below $z - \zeta_1$). This depth shift corresponds to half the lock-in depth. In the lower part only reversed magnetisations will contribute to the total remanence.

The model relies on the following assumptions:

1. Each magnetic grain is either locked or free to align its magnetic moment parallel to the external field. The lock-in function $\lambda$ depends only on

\[
M(z) = \int_{\zeta=0}^{z} \frac{\partial \lambda_{PDRM}(\zeta, c(z))}{\partial \zeta} H(z - \zeta) \, d\zeta
\] (6.3)
lithology $c(z)$ and burial depth $\zeta$, but not on the strength of the magnetic field ($\lambda = \lambda(\zeta, c(z))$. The dependence on field strength has not been considered because relevant experimental data are missing.

2. The lock-in rate is constant (linear model)
3. The remanence acquisition process is simply post-detrital.
4. The magnetic mineral assemblage in every horizon remains unchanged in concentration, composition and grain-size distribution while lock-in is taking place.

6.4.2 The linear detrital-pedogenic lock-in model

Since loess L8 at Lingtai contains both detrital and pedogenic magnetic minerals, a new approach to the lock-in problem is required. Two remanence acquisition processes are incorporated in the definition of the lock-in function $\lambda$ (Figure 6.5). Below a certain depth of loess sedimentation ($d_0$), the detrital population starts blocking with a certain lock-in rate. The PDRM is blocked

![Figure 6.5: General lock-in function $\lambda_{PDRM+CRM}(\zeta, e)$ and percentage of the detrital component for the detrital-pedogenic remanence model, which assumes that loess is being deposited and subsequently altered. Until the initial lock-in depth $d_0$ is reached, no magnetisation is acquired. The PDRM in the loess is locked completely after $d_1$ has been passed. The loess alteration starts at $d_2$ and new magnetic minerals can acquire a CRM, which is totally locked at $d_3$. The ratio PDRM/CRM in a stratigraphic layer is controlled by the percentage of the detrital component $e$ (vertical dashed line).]
entirely at depth $d_1$. Meanwhile chemical weathering begins and produces a new secondary magnetic mineral population, which starts to acquire a CRM at $d_2$. The second lock-in phase is due to the chemical remanent magnetisation (CRM), which is completed at $d_3$. The lock-in rate for both processes may be different. The corresponding two-stage lock-in function $\lambda$ can be expressed in the following way:

$$\lambda_{PDRM+CRM} = \lambda(\zeta, e(z)) = \begin{cases} 
0 & \text{for } \zeta < d_0 \\
e(z) \frac{\zeta - d_0}{d_1 - d_0} & \text{for } d_0 \leq \zeta < d_1 \\
e(z) & \text{for } d_1 \leq \zeta < d_2 \\
e(z) + (1 - e(z)) \frac{\zeta - d_2}{d_3 - d_2} & \text{for } d_2 \leq \zeta < d_3 \\
1 & \text{for } \zeta \geq d_3
\end{cases} \quad (6.4)$$

Note that the depth ranges of the two lock-in zones for PDRM and CRM are taken as constants throughout the model space. The newly introduced PDRM/CRM ratio $e(z)$ which describes lithogenic change, quantifies the relative

**Figure 6.6:** Schematic lock-in process of the alteration (detrital and pedogenic) remanence. The remanence is composed of contributions (arrows) from different discrete lock-in zones separated by the lock-in isochrons (dashed). Equally shaded arrows indicate their affiliation to the respective lock-in zone. A change in magnetic polarity will be observed if more than 50% of the total magnetic moments are blocked in the new field direction.
contribution of the PDRM (loess) fraction to the total NRM. Accordingly, 100-
$e(z)$ represents the NRM percentage that is due CRM (palaeosol). The
coefficient $e(z)$ can vary between 0 and 100 %, i.e. between the two theoretical
lock-in model end-members of 'pure loess' and 'pure palaeosol'. A sketch of the
two-component lock-in process during a geomagnetic field reversal is given in

**Figure 6.7:** Modelled detrital-pedogenic remanence for different constant detrital contributions. Left: The detrital contribution is constant over the whole model space at five different values (0, 25, 50, 75, 100%). Middle: A geomagnetic field reversal is modelled as a tanh function centred at 4 m depth. Right: The modelled remanence acquisition is divided into two parts. The part from 0 to about 3.5 m describes the lock-in process from the surface. Since the field polarity does not change during this most recent lock-in, the lock-in function is not convolved with the derivative of the geomagnetic field. This part of the remanence simply reflects the lock-in function (first term of equation (6.6)). If the polarity of the field changes during lock-in, the lock-in function is convolved with the derivative of the field. Therefore a distorted image of the lock-in process is obtained (it is convolved). The value of the plateau near 5 – 6 m depends on the lithogenic ratio $e$. Small lithogenic variations can easily cause multiple polarity flips of the magnetisation, if both mineral components (detrital and chemical) are nearly equally represented.
Figure 6.6. The remanence contribution of each lock-in zone to the total remanence depends on the variation of \( e \) with depth.

The strength of the reversing geomagnetic magnetic field \( H \) will be simulated by a hyperbolic tangent function ranging from \(-1\) to \(1\):

\[
H(z) = -\tanh\left[\frac{c_2 + c_1}{c_2 - c_1}\left(z - \frac{c_2 + c_1}{2}\right)\right]
\]

(6.5)

The constants \( c_1 \) and \( c_2 \) represent the end and the beginning of the reversal, respectively. The application of the linear detrital-pedogenic lock-in model to different artificial PDRM/CRM ratios is shown in Figure 6.7, \( e(z) \) being assumed constant. Nearly equal contributions of PDRM and CRM constitute an extremely sensitive balance for polarity changes and result easily in recording of multiple polarity flips despite a simple step-function reversal.

The model relies on the following assumptions: Each magnetic grain is either locked or free to align its magnetic moment parallel to the external field. The lock-in function \( \lambda \) depends only on lithology \( e(z) \) and burial depth \( \zeta \); \( \lambda = \lambda(\zeta, e(z)) \). The lock-in rate for each process is constant (linear model).

**Estimation of the model parameters**

The model assumes that detrital lock-in starts when a loess layer of thickness \( d_0 \) has accumulated. This may be due to “loessification” which is known to take place between about 0.4 and 1 m (Assallay et al., 1998); these two values are taken to be appropriate estimates for \( d_0 \) and \( d_1 \), respectively. It is assumed that pedogenic remanence carriers may start precipitating already during loess deposition. A pedogenic CRM in those mineral phases may begin blocking shortly after the detrital population has been locked. A small depth difference between \( d_1 \) and \( d_2 \) of only 0.20 m was chosen at first, corresponding to a time interval of ~2500 years and a value of 1.2 m for \( d_2 \). A value of 3.2 m was found to be appropriate for \( d_3 \). This lock-in depth corresponds to half of the MBB shift on the CLP (see Heslop et al., 2000; Zhou and Shackleton, 1999).

The function \( e(z) \) was estimated from detailed anhysteretic remanent magnetisation (ARM) measurements. The ARM is supposed to respond to the mineral populations that carry the detrital and pedogenic NRM components.
When an ARM (maximum AF of 150 mT, DC bias field of 0.05 mT) was subjected to stepwise AF demagnetisation, different lithologies were found to exhibit two distinct coercivity populations (Figure 6.8). The derivative of the demagnetisation curve of palaeosol S1 is dominated by a single maximum located at 21 mT, whereas the pristine loess L4 has a major peak at 44 mT in addition to the much less developed pedogenic component. The maxima of the gradient curves of ten previously investigated samples (not shown here) vary within intervals of 1.3 mT and 1.5 mT for the low and high coercivity samples, respectively. This method therefore provides robust estimates of the lithogenic ratio function $e(z)$. For the 137 samples at Lingtai between depths of 60.32 and 63.38 m, we chose the central difference quotients at 21 and 44 mT were chosen to represent the pedogenic and detrital remanences, respectively. The ARM coercivity spectra for the endmembers clearly show two separate peaks.

When an ARM (maximum AF of 150 mT, DC bias field of 0.05 mT) was subjected to stepwise AF demagnetisation, different lithologies were found to exhibit two distinct coercivity populations (Figure 6.8). The derivative of the demagnetisation curve of palaeosol S1 is dominated by a single maximum located at 21 mT, whereas the pristine loess L4 has a major peak at 44 mT in addition to the much less developed pedogenic component. The maxima of the gradient curves of ten previously investigated samples (not shown here) vary within intervals of 1.3 mT and 1.5 mT for the low and high coercivity samples, respectively. This method therefore provides robust estimates of the lithogenic ratio function $e(z)$. For the 137 samples at Lingtai between depths of 60.32 and 63.38 m, we chose the central difference quotients (typically on the order of $10^{-6}$ A/m/mT) around the two peak values as a measure of the contribution of detrital and pedogenic components. The lithogenic ratio was then calculated as:

$$e(z) = \frac{\Delta ARM_{44 mT}(z)}{\Delta ARM_{21 mT}(z)} \times 100\%$$  \hspace{1cm} (6.6)
Repeat measurements of selected samples indicate that the error in \( e(z) \) is between 1 and 3 \%.

**Sensitivity Tests**

Before applying the detrital-pedogenic model to the loess/palaeosol sediments of Lingtai using the measured ARM properties, it was first tested using arbitrary \( e(z) \) functions. The lithogenic ratio was assumed to vary sinusoidally between 0 and 100 \% every 0.2 m over a model space of 8 m. A field change from reversed to normal polarity is supposed to start at 4.15 m and to last for 0.4 m.

First the lock-in function is depicted at a level where sedimentation has stopped at 0 m (Figure 6.9a). The detrital remanence (peaking in cycles with lighter shading) starts to build up from zero once \( d_0 \) has been exceeded. No CRM (peaking in cycles with darker shading) forms at this stage. Below \( d_2 \) (at 1.20 m), the CRM also starts to lock-in and the PDRM is locked completely with maximum values of 1. The total remanence signal is constant below \( d_3 \) (3.20 m).

Of course, the lock-in function moves progressively upward with ongoing sedimentation. We start out in the reversed field. Hence, at depth, both components are blocked with reversed polarity. When the field reverses, the lock-in function comes into play and determines the proportion of normally and reversely locked remanence components, respectively. Since the pedogenic component is locked later than the detrital component, it acquires an increasingly normal component that is displaced downward (i.e. it occurs apparently earlier than it should). This holds true also for the detrital component at a later stage or nearer the actual reversal boundary. The convolution of the lock-in function with the field behaviour leads to the recording of seven apparent polarity flips. The normal polarity remanence is completely blocked 0.4 m below the end of the field reversal.

An increase in the period of lithogenic ratio \( e(z) \) fluctuations will reduce the number of polarity changes. If the period exceeds 0.86 m (as found by trial and error), only a single polarity change will be recorded Figure 9b. Small changes in the PDRM/CRM ratio occurring near a field polarity change can have a strong effect on the modelled NRM polarity pattern. Model calculations show
Figure 6.9: Test of the detrital-pedogenic lock-in model with artificial lithologies. a) The lithogenic ratio $e(z)$ varies sinusoidally between 0 and 100 % every 0.2 m. The modelled detrito-pedogenic remanence changes polarity seven times. b) Sensitivity of the model with respect to small lithogenic changes. Only pedogenic CRM controls the magnetisation with one exception where 25% of detrital material is added over a small thickness of 0.02 m at 6.06 m depth. Since this happens near the polarity transition of the CRM, it is enough to cause the polarity of the magnetisation to flip once more into reversed polarity.
that a 2 cm thick layer with a detrital contribution of 25% at a critical depth in an entirely pedogenic lithology may create an additional apparent polarity zone.

Model for the Lingtai MBB

Equation (6.3) describes the lock-in process over the whole model space. The lock-in process at the present surface is not of interest for modelling the MBB at Lingtai. Thus, we can transform e.g. (6.3) into:

\[
M(z) = \lambda_{PDRM+CRM}(z, e(z)) + \int_{\zeta=0}^{z} \lambda_{PDRM+CRM}(\zeta, e(z)) \frac{\partial H(z - \zeta)}{\partial \zeta} d\zeta
\]

(6.7)

The first term of (6.7) describes the lock-in process at the model surface when \(\zeta = z\). As this term is not convolved with the magnetic field, it is identical to the lock-in function. The second term represents the convolution of the lock-in function with the derivative of the magnetic field (see Figure 6.7). After lock-in depth \(d_3\) has been reached, the modelled remanence is constant at the maximum value (=1), thus equation (6.7) changes to:

\[
M(z) = 1 + \int_{\zeta=0}^{z} \lambda_{PDRM+CRM}(\zeta, e(z)) \frac{\partial H(z - \zeta)}{\partial \zeta} d\zeta
\]

(6.8)

The detrital-pedogenic model assumes that: (1) the correct stratigraphic position of the MBB is in the lower part of S7 as discussed above, and (2) the MBB magnetic field change takes place over a depth interval corresponding to \(~5000\) years at Lingtai. The value of 5000 years for the MBB reversal was chosen as a compromise between the estimated theoretical and observed geomagnetic reversal durations which range between 1000 and 8000 years (see Merrill and McFadden, 1999). Hence, the constants \(c_1\) and \(c_2\) in equation (6.5) are taken as 59.75 m and 60.15 m, respectively.

Both remanence acquisition processes (PDRM and CRM) will not necessarily contribute equally to the natural remanence. With respect to this, the NRM and ARM of two end-member samples were compared. The samples originate from the loess horizon L4 (at 30.42 m) which is dominated by the detrital component and the palaeosol S5 (at 40.62 m) which is dominated by the pedogenic
component. Using an average sedimentation rate of 7.9 cm/kyr during the Brunhes chron, ages of 385 ka and 514 ka are obtained for these two depths, respectively. According to Guyodo and Valet (1999), the strength of the geomagnetic field was approximately the same at both these times, within the uncertainties of their determinations. The following ratio can be regarded as an intensity indicator of the characteristic component of the NRM:

\[
\frac{NRM_{20mT} - NRM_{50mT}}{ARM_{20mT} - ARM_{50mT}}. \tag{6.9}
\]

The secondary magnetisation overprint is usually removed at 20 mT and above 50 mT the NRM becomes unstable (in loess samples). Both detrital and pedogenic components contribute to this magnetisation window of the characteristic NRM. If the intensity of the magnetic field is constant, any differences in this ratio are due to different efficiencies of remanence.

<table>
<thead>
<tr>
<th>demagnetising AF field [mT]</th>
<th>NRM</th>
<th>ARM</th>
<th>(\Delta NRM/\Delta ARM)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>20</td>
<td>50</td>
<td>20</td>
</tr>
<tr>
<td><strong>Loess L4</strong> ~ 385 ka</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>depth [m]</td>
<td>sample</td>
<td></td>
<td></td>
</tr>
<tr>
<td>30.36</td>
<td>A1518</td>
<td>7.13E-05</td>
<td>4.68E-05</td>
</tr>
<tr>
<td>30.38</td>
<td>A1519</td>
<td>6.11E-05</td>
<td>3.82E-05</td>
</tr>
<tr>
<td>30.42</td>
<td>A1521</td>
<td>7.00E-05</td>
<td>4.47E-05</td>
</tr>
<tr>
<td>30.44</td>
<td>A1522</td>
<td>9.14E-05</td>
<td>6.00E-05</td>
</tr>
<tr>
<td>30.46</td>
<td>A1523</td>
<td>8.61E-05</td>
<td>5.83E-05</td>
</tr>
<tr>
<td>mean</td>
<td>0.174</td>
<td>0.028</td>
<td>16.0</td>
</tr>
<tr>
<td>standard deviation</td>
<td></td>
<td>0.007</td>
<td>6.7</td>
</tr>
<tr>
<td>rel. standard deviation</td>
<td></td>
<td>16.0</td>
<td>6.7</td>
</tr>
</tbody>
</table>

| **Palaeosol S5** ~ 514 ka |     |     |     |     |                |
| depth [m] | sample |         |         |         |         |         |         |         |         |         |         |         |         |         |         |
| 40.56   | A2028  | 3.42E-04 | 8.16E-05 | 2.71E-03 | 1.94E-04 | 0.103   |
| 40.58   | A2029  | 3.98E-04 | 8.99E-05 | 2.99E-03 | 1.83E-04 | 0.110   |
| 40.62   | A2031  | 2.76E-04 | 6.30E-05 | 2.09E-03 | 1.56E-04 | 0.110   |
| 40.64   | A2032  | 2.69E-04 | 6.82E-05 | 2.30E-03 | 1.52E-04 | 0.093   |
| 40.70   | A2035  | 3.87E-04 | 9.37E-05 | 2.92E-03 | 2.19E-04 | 0.109   |
| mean    | 0.105  | 0.007 | 6.7 |                |
| standard deviation |                | 0.007 | 6.7 |                |
| rel. standard deviation |                | 16.0 | 6.7 |                |

**Table 6.2:** The samples originate from the most pristine loess L4 and well the developed palaeosol S5, both from Lingtai. The magnetic units are given in Am². The peak AF for the acquired ARM was 150 mT and the bias field 0.05 mT.
acquisition. The ratio is about \((0.174 \pm 0.028)\) in the loess sample and \((0.105 \pm 0.007)\) in the palaeosol (see Table 6.2) which implies that the NRM is more efficiently (~1.7 times) acquired by PDRM acquisition processes than by CRM acquisition. The lithogenic parameter \(e\) has therefore been corrected for NRM acquisition and recalculated:

\[
e_{\text{corr}} = \frac{0.174e}{0.174 + 0.105(100 - e)} \cdot 100\%.
\] (6.10)

The variation of the PDRM/CRM ratio \(e_{\text{corr}} (z)\) from S7 to S8, as given by ARM measurements e.g. (6.6), has been plotted in Figure 6.10. Higher detrital percentages are reached in the upper part of loess L8. The distinct decrease toward the pedogenically pre-dominated palaeosol shows that the ARM signal is much more sensitive to loess alteration than the magnetic parameters shown in Figure 6.2.

The modelled polarity pattern shows some common and some differing features with the observed NRM polarity (Figure 6.10a). Three polarity changes occur in L8 between 60.86 and 61.26 m. This is not in agreement with the thermal demagnetisation results. Hence, the initially assumed lock-in depths have been slightly modified to improve agreement with the observed data. The best-fitting lock-in depths are \(d_0 = 0.4\) m, \(d_1 = 1.60\) m, \(d_2 = 1.70\) m and \(d_3 = 3.2\) m (Figure 6.10b). After model recalculation, seven polarity changes between 61.48 m and 61.80 m occur. Now, the difference between the midpoints of both multiple magnetisation polarity intervals is only 0.03 m.

**Figure 6.10 (following page):** The detrital-pedogenic linear model applied to the loess/palaeosol sediments at Lingtai (central CLP). The changing pedogenic and detrital components cause multiple polarity changes in the modelled remanence. The resolution of the model is 0.02 m. **a)** The assumed lock-in depths (see text) do not lead to a good coherence between observed and modelled magnetisation polarity flips. **b)** Small changes of the lock-in depths lead to an improved agreement with the observed data. The lock-in depths imply nearly equal lock-in rates for the detrital and pedogenic lock-in process. Note that the lithogenic ratio \(e_{\text{corr}}\) in loess and palaeosol never accomplishes the values of the ‘pure’ endmembers.
**a) Detrital–pedogenic remanence – linear model**

<table>
<thead>
<tr>
<th>$\chi$</th>
<th>Detrital component</th>
<th>Assumed field</th>
<th>Total remanence</th>
<th>Modelled polarity</th>
<th>Observed polarity</th>
</tr>
</thead>
<tbody>
<tr>
<td>[$10^{-8} \text{ m}^3/\text{kg}$]</td>
<td>e$_{corr}$ [%]</td>
<td>behaviour</td>
<td></td>
<td>(thermal)</td>
<td>(thermal)</td>
</tr>
</tbody>
</table>

lock-in depths

d$_0$ = 0.40 m, d$_1$ = 1.00 m, d$_2$ = 1.20 m, d$_3$ = 3.20 m

**b) Detrital–pedogenic remanence – linear model**

<table>
<thead>
<tr>
<th>$\chi$</th>
<th>Detrital component</th>
<th>Assumed field</th>
<th>Total remanence</th>
<th>Modelled polarity</th>
<th>Observed polarity</th>
</tr>
</thead>
<tbody>
<tr>
<td>[$10^{-8} \text{ m}^3/\text{kg}$]</td>
<td>e$_{corr}$ [%]</td>
<td>behaviour</td>
<td></td>
<td>(thermal)</td>
<td>(thermal)</td>
</tr>
</tbody>
</table>

lock-in depths

d$_0$ = 0.40 m, d$_1$ = 1.60 m, d$_2$ = 1.70 m, d$_3$ = 3.20 m
6.4.3 The exponential detrital-pedogenic lock-in model

The linear lock-in model is a first-order approximation for remanence acquisition processes on the CLP. Several authors have suggested that exponential lock-in rates may be much closer to reality (Verosub, 1977; Hamano, 1980; Denham and Chave, 1982). In the following, an attempt is made to consider some physical processes for the lock-in process in order to approach a more realistic model.

The PDRM lock-in process in loess is mainly controlled by the pore size $p$. At shallow depth, the pore size is large enough to allow rotational movement of magnetic particles (or of those to which they are attached). The overburden due to continuing sedimentation causes linearly increasing pressure and reduced pore size with depth. Hamano (1980) used the void ratio ($= \text{volume of air and liquids/volume of solids}$) as a measure of pressure increasing with depth. In Chinese loess, the void ratio decreases due to compression of the pores, but grain size remains constant (Suzuki and Matsukura, 1992). Therefore, the relationship given by Hamano (1980) may be generalised by the following function. The mean pore size $\mu_d(\zeta)$ decreases with increasing burial depth $\zeta$:

$$\mu_d(\zeta) = p_t + (p_i - p_t) \cdot e^{-c_d \zeta}, \quad (6.11)$$

where $c_d$ is a constant (the subscript $d$ referring to the detrital process). $p_i$ and $p_t$ represent initial and terminal pore size, respectively. With increasing burial, the fraction of physically blocked moments is then:

$$\lambda_{PDRM}(\zeta) = \int_0^{p_b} \frac{1}{\sigma_d \sqrt{2\pi}} e^{-\frac{(\ln p - \mu_d(\zeta))^2}{2\sigma_d^2}} dp. \quad (6.12)$$

A log-normal pore size distribution moves through the critical pore size $p_b$ at which the particles are mechanically fixed. $\sigma_d$ denotes the standard deviation. Our model calculations use values for $p_i$ and $p_t$ from pore size distributions of sandy Malan loess given by Suzuki and Matsukura (1992). The pore size histogram of this loess exhibits two peaks, at 10 $\mu$m and 0.05 $\mu$m. The unknown parameters of blocking pore size $p_b$, constant $c_d$ and $\sigma_d$ were chosen in such a
manner that the detrital lock-in function starts to significantly deviate from zero at about 0.4 m and to reach its maximum at around 1.60 m as discussed above. In this way, values of 0.5 \( \mu \) for \( p_b \), 3.4 for \( c_d \) and 0.6 for \( \sigma_d \) were obtained (Figure 6.11).

The CRM acquisition process strongly depends on the volume of the remanence carriers. Grains below the blocking volume \( V_b \) cannot retain a remanent magnetisation. At the initial stage of pedogenesis, a log-normal distribution of newly forming magnetic grains has its mean \( \mu_p(\zeta) \) at \( V_0 \), well below \( V_b \). With ongoing time the magnetic particles grow. The distribution crosses \( V_b \) and the CRM becomes blocked. We assume that the process of crystallisation obeys the same mathematical rules as its reverse process, dissolution. The curve of surface dissolution of goethite, for instance, has a sigmoidal shape (Stucki et al., 1988) due to limitation of the dissolution rate.

Figure 6.11: a) Log-Gaussian distribution of pore sizes. Below a certain pore size, the grains are mechanically fixed and the magnetisation is blocked (left, shaded area). b) Mean of the distribution as a function of burial depth; initial value: \( p_i \). Below a certain depth, the terminal pore size \( p_t \) is reached. c) The percentage of blocked grains calculated by integration from 0 to the blocking pore size \( p_b \). The magnetisation starts blocking at \( d_0 = 0.70 \) m and is locked at \( d_1 = 1.76 \) m. (The lock-in depths correspond to the best-fit parameters of the exponential model, see Figure 6.13b)
This approach is comparable to surface-controlled chemical processes (Cornell and Schwertmann, 1996; van Oorschot, 2001). Thus, the volume change per unit time is assumed to be proportional to the volume $V$. It is further assumed that the concentration of iron ions in the fluid phase of the soil decreases with increasing crystallisation. This fixes the limit of the crystallisation process described by the term $\left(1 - \frac{V}{V_e}\right)$ with $V_e$ denoting the end volume:

\[
\chi_{ARM} \propto \frac{dV}{dt} = kV
\]

where $k$ is a constant. The term $\frac{dV}{dt}$ represents the volume change per unit time and $\chi_{ARM}$ is the ARM susceptibility. The proportionality constant $k$ is determined by the specific conditions and processes involved in the crystallisation. This relationship allows for the calculation of the volume change and the consequent changes in the ARM susceptibility, providing insights into the mineralogical evolution of the soil material with depth.
\[
\frac{dV}{dt} = cV \left( 1 - \frac{V}{V_e} \right).
\]  

(6.13)

The process starts at time \( t_0 \) with an initial volume \( V_0: V(t_0) = V_0 \). The solution of the differential equation is then (Figure 6.12c):

\[
V(t) = \frac{V_e}{1 + \left( \frac{V_e}{V_0} - 1 \right) e^{c_p t_0 + c_p t}},
\]

(6.14)

with the constant \( c_p \) representing the (unknown) growth rate. Two problems arise with equation (6.13). First, there is no useful information about the time dependence of the crystallisation processes of magnetite in soils. Second, our calculations are performed in the depth domain and not in time. Therefore, we arbitrarily adapt equation (6.13) to the burial depth \( \zeta \):

\[
V(\zeta) = \frac{V_e}{1 + \left( \frac{V_e}{V_0} - 1 \right) e^{c_p t_0 + c_p \zeta}}.
\]

(6.15)

Starting out from a nucleation grain volume of \( V_0 = 1 \times 10^{-25} \text{ m}^3 \) (at depth \( \zeta_0 \)), the grain growth is arbitrarily assumed not to exceed the pseudo-single domain range (PSD) and, hence, the grain diameter will not exceed 0.09 \( \mu \text{m} \) \( (V_e = 3.82 \times 10^{-22} \text{ m}^3) \). According to Moon and Merrill (1985), this is the upper limit for single domain magnetite. If a magnetic grain exceeds the stable single domain (SSD) size, the particle contains more than one domain and the magnetisation markedly decreases (Figure 6.12a). The SSD grain size has been estimated for magnetite to lie between 0.05 to 0.06 \( \mu \text{m} \), which corresponds to a volume of \( V_{SSD} \approx 1.13 \times 10^{-22} \text{ m}^3 \) (Dunlop, 1973; Argyle and Dunlop, 1984).

The grain size dependence of the magnetisation has to be taken into account for the blocking of pedogenic grains. ARM versus grain size data (Dunlop and Xu, 1993; Egli and Lowrie, 2002) have been translated into volumetric data (assuming spherical shape) and fitted with power-law functions (Figure 6.12a). The resulting function acts as a weighting factor for the pedogenic lock-in process:
\[
\chi_{ARM}(V) = \begin{cases} 
10^{11.5565} \cdot \left(\frac{6V}{\pi}\right)^{\frac{1.931}{3}} & V < V_{SSD} \\
10^{-5.6689} \cdot \left(\frac{6V}{\pi}\right)^{-\frac{0.455}{3}} & V > V_{SSD}
\end{cases}
\] (6.16)

By analogy with equation (6.12) the proportion of pedogenically blocked grains (parameters with subscript \(p\)) is given by (Figure 6.12d):

\[
\chi_{CRM}(\zeta) = \int_{V_b}^{\infty} \chi_{ARM}(V) \cdot \frac{1}{\sigma_p \sqrt{2\pi} V} e^{-\frac{(\ln V - \mu_p(\zeta))^2}{2\sigma_p^2}} dV.
\] (6.17)

The blocking volume \(V_b\) is assumed to be 8.18x10^{-24} m^3. This value implies a grain size of 0.025 \(\mu\)m, which is the lower limit grain size for SD magnetite (McNab et al., 1968).

Since sedimentation continues during interglacial stages on the central CLP, burial depth is related to deposition time. Volume growth as a function of increasing burial is shown in Figure 6.12c. The unknown parameters \(\sigma_p, c_p, \zeta_0\) were adapted in order to fix the pedogenic lock-in function (Figure 6.12d) between the best-fit lock-in depths \(d_2\) and \(d_3\) derived from the linear model (cf. Figure 6.10b).

Calculating the remanence with this non-linear lock-in function leads to more multiple polarity flips (Figure 6.13a) compared to the linear model (Figure 6.10b). The non-linear lock-in model, however, does not fit the observed data particularly well and significantly deviates from the best-fit of the linear model (Figure 6.10b). Changing \(c_d, \sigma_d, c_p\) and \(\zeta_0\) to 2.55, 0.4, 5.75 m^2 and 1.12 m, respectively, yields best-fit lock-in depths of \(d_0 = 0.7\) m, \(d_1 = 1.76\) m, \(d_2 = 1.82\) m and \(d_3 = 3.2\) m. The results of the exponential lock-in model now become similar to the linear model (Figure 6.13b).

These lock-in depths differ considerably from those initially assumed. In view of their uncertainty, it is not worthwhile discussing their accuracy. The linear and the exponential lock-in models both show that small variations of the lithogenic properties can affect the polarity of the magnetisation during a reversal. As pedogenic action on the moist central CLP can start as soon as the loess is deposited, the exponential model may be more realistic.
a) Detrital–pedogenic remanence – exponential model

\[ \chi \]

<table>
<thead>
<tr>
<th>Depths [m]</th>
<th>Detrital component [10^{-8} m^3/kg]</th>
<th>Assumed field [e_{corr} %]</th>
<th>Total remanence</th>
<th>Modelled polarity</th>
<th>Observed polarity</th>
<th>VGP lat. [°]</th>
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</thead>
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lock-in depths
- \( d_0 = 0.40 \text{ m} \), \( d_1 = 1.60 \text{ m} \), \( d_2 = 1.70 \text{ m} \), \( d_3 = 3.20 \text{ m} \)

b) Detrital–pedogenic remanence – exponential model

\[ \chi \]

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<thead>
<tr>
<th>Depths [m]</th>
<th>Detrital component [10^{-8} m^3/kg]</th>
<th>Assumed field [e_{corr} %]</th>
<th>Total remanence</th>
<th>Modelled polarity</th>
<th>Observed polarity</th>
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lock-in depths
- \( d_0 = 0.70 \text{ m} \), \( d_1 = 1.76 \text{ m} \), \( d_2 = 1.82 \text{ m} \), \( d_3 = 3.20 \text{ m} \)
6.5 Discussion and Conclusions

Loess is mechanically unstable after deposition. Below a certain depth, it compacts into a more rigid sediment as the transformation of dust into loess, coupled with secondary calcification, builds up a microstructure. This structural change takes place between ~0.4 and ~1 m under specific loading and wetting conditions (Assallay et al., 1998), and, among other things, leads to the mechanical fixing of the magnetic grains. Our linear modelling results for the Lingtai section support the idea that the post-depositional remanent magnetisation is blocked over approximately this depth zone.

Under sufficiently moist and warm climate conditions, chemical weathering releases iron ions as a basis for the formation of secondary pedogenic iron minerals. Evans and Heller (1994) proposed the coexistence of two magnetic mineral components in loess/palaeosol sediments on the CLP: the primary component is a detrital assemblage of magnetic particles, which appears to be uniform across the entire loess plateau. The second component is an authigenic phase varying between sites and also from layer to layer in a section. The measurements and modelling presented here confirm this suggestion: a low coercivity mineral component (P) is enriched with increasing degree of pedogenesis, whereas the detrital component (D) remains rather constant.

Mineral-magnetic analysis suggests that post-depositional and (bio-) chemical recording processes occur at different stages of burial and therefore at different moments in time and in geomagnetic field history. The relative contribution of the two recording mechanisms has been estimated on the basis of differences in the ARM coercivity spectra. The two lock-in models considered in this paper take both, PDRM and CRM acquisition into account. Piecewise linear lock-in functions lead to a reasonable agreement with the observed data. The exponential model, which seems to be more relevant physically, yields essentially the same results.
Considering that low values and variations of the modelled remanence intensities allow for multiple polarity flips, one could argue that these signal variations are simply random noise. However, the occurrence of a low magnetisation intensity zone is an important property of the two component lock-in model if detrital and authigenic carriers contribute at nearly equal proportions (see zero remanence plateau of the 50% PDRM/CRM curve in Figure 6.7). Consequently, minor lithogenic variations can easily cause multiple polarity flips, which by no means are caused by rapid geomagnetic field changes. Although the intensity of the magnetic field may also be reduced, the low magnetisation zone is mostly due to the lithogenic properties of the sediment and the secondary formation of magnetic minerals. The actual data at Lingtai also exhibit such a zone of low NRM intensity during the MBB transition (Spassov et al., 2001).

A general problem of lock-in processes is the incorporation of depth and time dependent processes. The fixing of magnetic grains in loose sediments depends mainly on the pore size, which decreases with increasing pressure resulting from increasing burial. Therefore the depositional lock-in process depends on burial depth and not on burial time (Bleil and von Dobeneck, 1999). The change of the Earth’s magnetic field on the other hand depends on time, not on burial. Thus the calculated magnetisation is a function of both time and depth. Several authors simplify the calculations and perform model calculations in only the depth domain (Bleil and von Dobeneck, 1999; Hyodo, 1984). Calculations in the depth domain are legitimate only for depositional lock-in processes because the transformation from time to depth is linear. This may be not the case for CRM acquisition. The CRM lock-in process depends strongly on the grain volume, which is supposed to increase with time. The burial dependence is indirectly incorporated by assuming ongoing sedimentation of loess during soil development, which is the case in areas of high dust accumulation such as the CLP. Sediment accumulation during soil development and, thus, the indirect burial dependence of CRM acquisition is not well known.

With these precautions in mind, the following conclusions can be drawn from modelling the NRM lock-in process at Lingtai:

The linear detrital-pedogenic remanence model is able to explain the downward shift of the MBB in the loess/palaeosol sequence at Lingtai on the central CLP. A displacement from the stratigraphic expected level of 1.74 m (corresponding to a time delay of about 22 kyr) results from the model
considerations. The model further explains the observed occurrence of multiple polarity changes by small variations of closely balanced contributions of the detrital and pedogenic mineral components during the MBB polarity change. The good agreement of the modelled results with the observed data supports the hypothesis of two remanence acquisition processes to be involved in the loess sediments of the central CLP: post-depositional (PDRM) and chemical (CRM) remanent magnetisation.

The observed multiple polarity changes of the ChRM component are not features of the geomagnetic field during the MBB. They are caused by variable relative contributions of detrital and pedogenic magnetisation components during the reversal, which give rise to irregular polarity lock-in at the MBB. Hence, virtual geomagnetic pole paths during reversals from profiles of the central CLP are not representative of geomagnetic field behaviour.

The actual physical deposition processes are most probably not of a simple linear nature. Therefore an attempt has been made to explain the displaced and complicated MBB by a non-linear lock-in model. This model also explains the downward shift of the MBB and the multiple polarity changes. It is in better agreement with the observations than the linear model. Further fundamental data such as blocking pore size, rate of pedogenic grain growth and lock-in depths are needed to obtain better information about the model parameters. The acquisition process of the NRM also needs to be studied. In addition, the simplistic function of the geomagnetic field during the M/B reversal may be replaced with more realistic data for the Earth’s magnetic field as given for instance by the SINT800 palaeointensity stack (Guyodo and Valet, 1999) or by reversal simulations (Coe et al., 2000).

The recognition of a delayed Matuyama-Brunhes boundary in loess/palaeosol sequences on the Chinese Loess Plateau solves their conflict with palaeoclimatic records from the marine realm. Hence the Chinese palaeosol S7 almost certainly corresponds to the marine oxygen isotope stage 19, as postulated by Heller et al. (1987), Zhou and Shackleton (1999) and Heslop et al. (2000).

The linear detrital-pedogenic lock-in model is proposed as a first step in solving delayed remanence acquisition processes in sediments with complex recording histories. It yields promising results in the loess/palaeosol sequences of the central CLP. The model can be tested using other well-defined reversal boundaries and other loess lithologies, for instance, on the western CLP. Further
experiments concerning the dynamics of CRM acquisition have to be performed to improve mathematical simulations.

Bleil and von Dobeneck (1999) already demonstrated for marine sediments that pseudo-records of palaeointensities can be obtained just by varying the lithology at low to intermediate sedimentation rates (1 - 4 cm/kyr). On the central CLP, two or more magnetic mineral populations with variable relative contributions almost always coexist. Their complex interaction and blocking will hamper or even prohibit useful information on palaeointensity changes, palaeosecular variation and polarity transition features during major reversals and/or excursions.
Part III

Conclusions
7. Conclusions and future research
7.1 Magnetisation and magnetic mineralogy in Chinese loess

The present study has been aimed at explaining and scrutinising the discrepancies between litho- and magnetostratigraphy between marine and continental sediments. In order to reach this goal, the directional variations around the Matuyama/Brunhes boundary (MBB) in the loess/palaeosol section at Lingtai (central Chinese Loess Plateau) have been studied in detail.

Inconsistent and rather erratic directional paths of the virtual geomagnetic pole (VGP) derived from characteristic NRM directions have been observed in the Lingtai loess/palaeosol record. It is suggested that they do not represent details of the MBB transitional geomagnetic field behaviour. The east Asian VGP reversal paths of the Matuyama-Brunhes boundary observed in different environments (continental loess and marine volcaniclastics) do not follow proposed phenomenological geomagnetic reversal models. The extremely low NRM_{20mT}/χ and NRM_{20mT}/M_{rs} ratios, the high MAD values, the variable ChRM polarity patterns and the multiple polarity changes are interpreted as reflecting complex magnetisation lock-in processes in Lingtai. They incorporate variable detrital and chemical magnetisation components, both in palaeosols and in loess layers. Assuming a constant sedimentation rate, these processes have been acting for up to 22000 yr after deposition as inferred from careful stratigraphic correlation (Heller et al., 1987; Zhou and Shackleton, 1999) and astronomical tuning (Heslop et al., 2000).

The largely delayed magnetisation lock-in limits the value of the loess on the central Chinese Loess Plateau as high fidelity recorder of geomagnetic field transitions and short-term excursions. Our study on the western Chinese Loess Plateau where higher sedimentation rates and drier climate are more favourable, failed for technical reasons. The borehole at Jingyuan jammed at a depth of 80 m, far above the expected MBB depth.

Magnetic methods in combination with grain size fractionation give evidence that the detrital and chemical magnetisations in the loess/palaeosol samples from Lingtai are carried by four coexisting coercivity populations. The two-component model of Evans and Heller (1994) has been extended to four components similar to the model of Eyre (1996). A mixing model based on the observed coercivity spectra of certain grain size fractions has been used to model IRM acquisition curves of loess/palaeosol samples. The coercivity
distributions of the grain size fractions do not depend on lithology only, but also on grain size. Each loess/palaeosol sample contains at least four different coercivity populations, which reside in detrital and pedogenic mineral phases. The contribution of the different coercivity components varies with lithology and has been quantified. It is argued that the detrital mineral components (hematite and maghemite) have not been destroyed during pedogenesis because they contribute equal amounts to the remanence of both, loesses and palaeosols. Hence, the iron for the new in-situ formed pedogenic (remanence carrying) mineral component must derive from clays and other ferruginous silicates. The iron of the clay-sized fraction has been mobilised during pedogenesis, but the iron content of loesses and palaeosols has been kept at a constant level. New ferromagnetic grains near and below the stable single domain boundary have grown within the same grain size fraction on translocated clays.

The magnetic mineral populations coexisting in loess/palaeosol sediments and the discrepant stratigraphic positions of the MBB in the marine and aeolian records have been the base for further developing the concept of delayed remanence acquisition. The linear detrito-pedogenic remanence lock-in model deployed in this study explains the apparent downward shift of the MBB in the loess/palaeosol sequence at Lingtai. The model copes with a displacement of 1.74 m from the stratigraphic level expected from palaeoclimate comparison with the marine magnetostratigraphy. This corresponds to a time delay of about 22 kyr. The model further explains the observed multiple polarity changes by variations of the relative contribution of the detrital and pedogenic mineral components in the interval of the MBB polarity change. The modelled results agree well with the observed data and support the hypothesis of postdepositional (PDRM) and chemical (CRM) remanence acquisition processes to be involved in the loess sediments of the central CLP. The multiple polarity changes of the characteristic NRM are not features of the geomagnetic field during the MBB. They are caused by variable relative contributions of detrital and pedogenic magnetisation components within the reversal interval, which give rise to irregular polarity lock-in around the MBB and possibly also during much of the loess deposition time. Hence virtual geomagnetic pole paths during reversals from profiles of the central CLP in general do not represent geomagnetic field behaviour. It is unlikely that short polarity excursions such as the Blake or Jamaica excursions have been recorded on the central CLP at all.
The actual physical deposition processes are most probably not of a simple linear nature. Therefore a non-linear lock-in model has been developed to explain the displaced and complicated MBB. This model also accounts for downward displacement of the MBB and multiple polarity changes. It agrees better with the observations than the linear model using best-fit parameters for the lock-in depth. The present model parameters still do not rely on solid experimental data. Blocking pore size and rate of pedogenic grain growth should be evaluated directly for the sections under investigation. In addition, the simplistic function used for the Matuyama/Brunhes geomagnetic reversal may be replaced with more realistic polarity transition data as given for instance by the SINT800 palaeointensity stack (Guyodo and Valet, 1999) or by reversal simulations (Coe et al., 2000).

A delayed Matuyama-Brunhes boundary in loess/palaeosol sequences on the Chinese Loess Plateau solves the conflict with records from the marine realm. Consequently, the Chinese palaeosol S7 corresponds to the marine oxygen isotope stage 19, as postulated by Heller et al. (1987), Zhou and Shackleton (1999) and Heslop et al. (2000).

7.2 Correlation between marine and continental sediments

The major goal – to explain the different stratigraphic positions of the Matuyama-Brunhes boundary (MBB) in the marine oxygen isotope records and the continental susceptibility records of the loess/palaeosol section of Lingtai – has been reached. The downward shift of the MBB in the loess/palaeosol sequence of Lingtai has been explained by delayed NRM acquisition. The corrected MBB position is now in the lower part of S7 and is dated at 778 ka (Tauxe et al., 1996). The time difference between observed and corrected MBB is about 22 kyr. The marine oxygen isotope and the continental susceptibility records with the corrected MBB can be correlated now, allowing to establish a time scale for loess/palaeosol sediments. Each major palaeosol horizon can be correlated with an odd numbered oxygen isotope stage (Figure 7.1) representing a warm and humid interglacial period.
Figure 7.1: Correlation between the marine oxygen isotope record at ODP site 677 (Shackleton and Hall, 1989; Shackleton et al., 1990) and the loess/palaeosol record at Lingtai (S=palaeosol, L=loess, Lca =carbonate concretion horizon within loess). The oxygen isotope record is tuned to an absolute timescale using the insolation curve of Berger (1987). The timescale of the loess record was obtained by matching begin and end of each odd numbered oxygen isotope stage with begin and end of each major palaeosol horizon. The dashed line in S7 represents the corrected Matuyama/Brunhes boundary (MBB), whilst the solid line in L8 is the observed one. The age of the MBB is 778 ka according to Tauxe et al. (1996).
It is further possible to calculate sedimentation rates for each major lithologic horizon using the loess/palaeosol time scale of Figure 7.1. Figure 7.2 shows large differences of the sedimentation rates in loesses and palaeosols. Sedimentation rates are on the average three times higher during glacial periods than during interglacials. It is further apparent that the accumulation of loess does not stop during the interglacials, it is just reduced. This is coherent with the present day interglacial situation when dust storms still accumulate loess over China.

![Figure 7.2: Sedimentation rates at Lingtai.](image)

**Figure 7.2:** Sedimentation rates at Lingtai. The time is based on the correlation given in Figure 7.1. Glacial loess layers (L) are characterised by much higher sedimentation rates than palaeosols (S) formed during interglacial periods. On the other hand, loess accumulation did not stop completely during interglacials.

### 7.3 A model of the loess/palaeosol ecosystem

The susceptibility record changes abruptly between loess and palaeosol lithology, especially in the younger part of the section (0 to 40 m, Figure 3.1a) where the contrast between loess and palaeosol is large. This is observed in many susceptibility profiles (Heller and Evans, 1995). Scheffer et al. (2001) proposed a simple model to explain drastic changes of ecosystem states. Smooth and gradual variations of the environmental conditions may alter its stability so that the system becomes more receptive to external influences. Small
perturbations at a point near instability are then enough to drop the system into
the other state. Following this model, loess/palaeosol sediments may be
characterised by two main stages (Table 7.1).

<table>
<thead>
<tr>
<th>Stage I - glacial</th>
<th>Stage V - interglacial</th>
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<tbody>
<tr>
<td>strong wind</td>
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<tr>
<td>high loess sedimentation</td>
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<tr>
<td>dry conditions</td>
<td>more precipitation</td>
</tr>
<tr>
<td>lower temperatures</td>
<td>higher temperatures</td>
</tr>
<tr>
<td>thin soil cover (xerosol)</td>
<td>thick soils (luvisol)</td>
</tr>
<tr>
<td>sparse vegetation</td>
<td>more vegetation</td>
</tr>
<tr>
<td>dominating winter monsoon</td>
<td>reduced winter monsoon</td>
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</table>

Table 7.1: Stable stages of the ecosystem loess/palaeosol.

The glacial period is predominated by strong winter monsoons. The Siberian
anticyclone may be stable also during summer times causing dry continental
summers. As a consequence much loess is blown onto the CLP resulting in high
sedimentation rates (see Figure 7.2). Under these conditions only sparse
vegetation is possible. During wet spring months some herbs (ephemeral plants)
are growing which wither during the hot summer. Strong winds may erode the
thin soil and cover the surface soon with new loess. The ground water table may
be deep and unreachable for the vegetation. When the orbitally controlled
insolation increases, the influence of winter and summer monsoon changes.
Now the moist summer monsoon is increasing and less loess is being
accumulated. The stable warm climate leads to the occurrence of perennial
vegetation, weathering of loess and soil development.

The abrupt periodical changes in loess/palaeosol sections can be described
with a folded ecosystem response curve (Scheffer et al., 2001) (Figure 7.3). At
stage I the ecosystem is in its first stable minimum (loess lithology). Increasing
insolation brings the ecosystem-state to the bifurcation-point F₂. At stage IV
only a slight incremental change is needed to bring the system beyond F₂ into
the alternative stable state. Its minimum (soil lithology) is reached at stage V.
Decreasing insolation leads to F₁. Further cooling and aridity bring the system to
the upper branch of stage II. Loess is being deposited again.
Deserts exist in China since 22 Myr (Guo et al. 2002b), but loess has not always been blown onto the Chinese Loess Plateau. The red clay underlying the loess/palaeosol sequences is also an aeolian deposit. Because of reduced wind activity its average sedimentation rate is with 3 cm/kyr much smaller than in the overlying loess/palaeosol sequence (Ding et al., 1999) and comparable to the that of palaeosols S1, S6, S7 and S8 (Figure 7.2). The red clay sequence can be regarded as a huge soil complex (Ding et al., 1999), in which aeolian material

**Figure 7.3:** Possible model of the loess/palaeosol ecosystem. Increasing insolation brings the ecosystem from state I to the bifurcation point F₂. A slight incremental change induces a sudden shift to a lower stable state (IV) with soil lithology. Decreasing insolation brings the system to F₁. Again a sudden change towards loess lithology can happen easily at F₁ (II). (Redrawn from Scheffer et al., 2001.)

### 7.4 The climatic shift at the Pliocene/Pleistocene boundary (~2.6 Ma)

Deserts exist in China since 22 Myr (Guo et al. 2002b), but loess has not always been blown onto the Chinese Loess Plateau. The red clay underlying the loess/palaeosol sequences is also an aeolian deposit. Because of reduced wind activity its average sedimentation rate is with 3 cm/kyr much smaller than in the overlying loess/palaeosol sequence (Ding et al., 1999) and comparable to the that of palaeosols S1, S6, S7 and S8 (Figure 7.2). The red clay sequence can be regarded as a huge soil complex (Ding et al., 1999), in which aeolian material
has been slowly added onto the soil surface during pedogenesis. The red clay is in that sense similar to palaeosols and contains also pedogenic and detrital components.

Loess accumulation started ~2.6 Ma ago. It initialised by the reorganisation of the pressure systems over Asia (Ding et al., 1992). According to general circulation models (GCM) the Tibetan Plateau plays an important role for the stability and position of the Siberian anticyclone (Mintz, 1965; Kasahara and Washington, 1969; Manabe and Terpstra, 1974, Zwiers, 1993; Meehl, 1994). Before the uplift of the Tibetan Plateau, the Siberian anticyclone was located much more south (30° N, 90° E) which led to much less wind over the source regions of the CLP (Manabe and Tepstra, 1974). When the main uplift started ~4.5 Ma ago (Zheng et al., 2000), the anticyclone moved to the northeast now affecting the source regions of the CLP. The ice ages were triggered by the closing of the Isthmus of Panama at the same time, which caused a new thermohaline circulation affecting predominantly the northern hemisphere. The beginning of loess sedimentation coincides with the massive appearance of ice-rafted debris in the northern Atlantic around 2.7 Ma.

7.5 Future research

We arrived now at the end of the alley and our journey is for the time being all over. We discovered many problems, tried to understand some things, and have found a few answers. Science is captured in a large palace with many corridors, doors and rooms. If we open one door, we come into another room with many other doors, which have to be opened. New questions arise and have to be answered, while still answering the old ones. In this sense an outlook for future research in Chinese loess will be attempted.

The topics are related to recent soils and palaeosol formation in different climates, loess alteration, acquisition of natural remanence and palaeoclimate.

- Pedogenic processes in recent soil and palaeosol profiles need to be further inspected. Chapter 5 concluded that detrital remanence carriers are not destroyed during pedogenesis and pedogenic iron oxides get the iron from detrital clays and ferruginous silicates. This statement has to be confirmed, because it relies on four samples only. Coercivity curves and thermomagnetic
experiments of grain size fractionated samples of a complete soil profile should be combined with X-ray diffraction, electron microscopy, iron chemistry ($\text{Fe}_\text{v}/\text{Fe}_\text{d}$) and iron isotope analysis in order to identify the genealogical pathways of iron during pedogenesis. The role of iron-assimilating bacteria for the production of pedogenic magnetite or maghemite in aerated soils is not clear at all. Laboratory experiments to grow bacteria under conditions, which are typical for soils should be carried out. The timescale for soil evolution and formation of magnetic minerals is presently unknown. Pristine loess might be put under suitable laboratory conditions to simulate weathering and pedogenesis, physical and chemical properties being measured during the experiment. The magnetic properties of recent soils have been correlated with present climate in order to establish “climofunctions” for reconstruction of the past climate. However, diagenetic processes are not considered in these models. Diagenesis during soil burial with oxidation of organic carbon, illitisation of smectites, dehydration and recrystallisation of iron and manganese oxyhydroxides and dolomitisation of pedogenic calcite may alter the characteristics of a soil. The diagenetic state of buried soils may be recognised from the comparison of recent soils and palaeosols. Present palaeoprecipitation models rely on magnetic parameters such as susceptibility but do not take diagenesis into account (Heller et al., 1993; Maher and Thompson, 1995). The susceptibility may increase when iron oxides recrystallise, which could lead to an overestimation of the palaeoprecipitation. Better knowledge of pedogenic and diagenetic processes is required to obtain more accurate information about the past climate from ancient soils.

- The consequences of loess alteration for the magnetic mineralogy in loess are just emerging right now. Chapter 5 presented evidence for a purely detrital ($\text{D}_1$) and an altered detrital ($\text{D}_2$) mineral component. The question arises if a loess, which is more pristine then Li-L4, has a $\text{D}_2$-component, too. If $\text{D}_2$ is also present in loesses on the western CLP, then the material in loess source regions (sand and gravel deserts) and unaltered recently deposited loess should also contain the $\text{D}_2$-component. The variable contribution of $\text{D}_2$ could then be interpreted in terms of different wind strengths. If $\text{D}_2$ is not present on the western CLP, its occurrence on the central CLP must depend on local climate. The coercivity spectra of a certain well recognisable loess layer,
could be analysed using material from places with different humidity and temperature.

- The lock-in model should be confirmed at different geomagnetic polarity boundaries in different loess sections. Its input parameters should be improved. Recent investigations Guo et al. (2002a) on the Chinese Loess Plateau; Oda et al. (2002) in Lake Baikal and this work (Figure 3.2) suggest that lithogenic properties affect the characteristic component of the natural remanent magnetisation. Different grain size fractions of Chinese loess/palaeosol samples contain different remanence carriers, which acquire their remanence at different times. Guo et al. (2002a) did not observe multiple polarity changes at the lower Jaramillo boundary in loess layer L12, where the median grain size is constant. They recorded them at the upper Jaramillo transition in S10/L10 where grain size varies due to lithological changes.

- The Mio-/Pliocene red clay is considered to be a soil complex with aeolian parent material. Therefore it should contain pedogenic and detrital remanence components, which may be identified by coercivity spectra analysis. The aeolian dust component may differ due to reduced wind strength although the source region may be the same as that of loess.

- The climate change at the Pliocene/Pleistocene boundary in China can possibly explained by the hysteretic model of Scheffer et al. (2001). To understand the impact of this major climate change on lithological and magnetic loess properties, detrital and pedogenic remanence carriers have to scrutinised more closely. The variation of the detrital components in the stratigraphic loess/palaeosol column should be studied by the rock magnetic techniques introduced and implemented in this study. In this way, palaeomagnetism can be developed into a valuable tool to understand better the past climates evolution as recorded in the huge aeolian sequences on the Chinese Loess Plateau.
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