



Rock-magnetic study of late Quaternary sediments from the eastern margin of the Tibetan Plateau

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博士論文

**Rock-magnetic study of late Quaternary sediments
from the eastern margin of the Tibetan Plateau**

(チベット高原東縁の第四紀堆積物の岩石磁気学的研究)

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Rock-magnetic study of late Quaternary
sediments from the eastern margin of the
Tibetan Plateau

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Contents

Acknowledgment	iii
Abstract	1
1 Introduction	3
2 Magnetic grain-size distribution of the enhanced component in the loess-paleosol sequences in the western Loess Plateau of China	7
2.1 Introduction	8
2.2 Samplings and laboratory methods	9
2.3 Results	11
2.4 Discussion	15
2.5 Conclusions	18
3 A record of late Holocene cooling event from Er-hai Lake, southwestern China	19
3.1 Introduction	20
3.2 Sampling and chronology	21
3.3 Rock-magnetic analysis	23

3.4	Palynological analysis	33
3.5	Discussion	37
3.6	Conclusions	41
4	Concluding remarks	42
	References	44
	Figure captions	55
	Figures	57

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Abstract

Rock-magnetic measurements were done on the two kinds of sediment from the eastern margin of the Tibetan Plateau. In both cases, the separation into two magnetic components is quite effective for accurate understanding of magnetic characteristics.

The background component and the enhanced component were imparted from the Chinese loess-paleosol samples from the Beiyuan section in Linxia City. The magnetic properties of the enhanced component suggest strong grain-size control, which is in contrast with variable grain-size distribution in loess/paleosol from the central Loess Plateau. Chemically formed magnetic grains is not a likely enhanced magnetic fraction in the case of the western Loess Plateau, because they show shifts in the grain-size distribution. The difference in climate between the western and central Loess Plateau may cause different origin and different preservation condition of the pedogenic magnetite.

Rock-magnetic and pollen analyses were done for the sediment cores from Erhai Lake, southwestern China. The change in magnetic mineralogy from greigite to magnetite and increase in portion of *Abies* suggest a transition to colder climate at about 3.5 ka. The period of the cooling event is consistent with a number of the

Neoglacial records from the Tibetan Plateau. The glacier advance on the Tibetan Plateau during the Neoglacial may have affected the climate in Yunnan Province.

Chapter 1

Introduction

The rock-magnetic properties of sediments are extensively investigated in order to reconstruct the Quaternary paleoclimatic and paleoenvironmental records. Environmental changes are often reflected on magnetic minerals as the dissolution and alteration of primary magnetic minerals, or as the formation of new magnetic minerals. They are detected as variations in composition and/or grain-size distribution of magnetic minerals using rock-magnetic techniques [Verosub and Roberts, 1995; Maher and Thompson, 1999a].

As a consequence of dissolution or formation of magnetic minerals, multiple magnetic components are likely to appear. In such cases, one magnetic parameter (for example, magnetic susceptibility alone) is insufficient to analyze complicated magnetic properties of sediments. However, many researchers tend to make use of single magnetic parameter and to regard it as a useful proxy of paleoclimate and paleoenvironment without understanding the controlling mechanism.

Here magnetic properties of two kinds of Quaternary sediments from the east-

ern margin of the Tibetan Plateau were studied in detail. Since the Tibetan Plateau occupies an important position in the Asian monsoon climate system as a barrier to wind and a heat source in summer, the climatic fluctuation around the Tibetan Plateau has been a subject of great interest. Especially, the uplift of the Tibetan Plateau [Kutzbach et al., 1989; Fang et al., 1999a] and the advance and regression of the ice sheet and glacier [Lehmkuhl and Haselein, 2000] have noticeable effect on climate in the Eastern Asia.

For the first target, the loess-paleosol sequences in the western Loess Plateau is taken up. In order to discuss the continental record of East Asian monsoon system, magnetic susceptibility of the loess-paleosol sequences is used as a representative proxy of paleorainfall. Earlier studies [Heller and Liu, 1982; Heller and Liu, 1984] has found that high magnetic susceptibility is recorded in paleosol, which is formed in the warm and humid condition during the interglacial and interstadial. The frequency-dependent magnetic susceptibility and high ARM susceptibility revealed that very fine magnetite (or maghemite) grains which were formed by some pedogenic action causes magnetic enhancement in paleosol [Heller and Evans, 1995; Maher and Thompson, 1999b].

Magnetic susceptibility as a proxy of paleorainfall is applied to various studies. Spatial variation of paleorainfall during interglacial and glacial periods in Pleistocene was reconstructed using a correlation between modern rainfall and magnetic susceptibility of modern soils [Maher et al., 1994]. The effect of the Tibetan uplift on the atmospheric circulation is also discussed on the basis of magnetic susceptibility [e.g. Fang et al., 1999a].

However, the detailed formation process of the very fine magnetite grains has

been open to argument. Some researchers [e.g. Maher and Thompson, 1999b] insist that it is inorganic precipitation around iron-reducing bacteria based on the result in the central Loess Plateau, whereas some [e.g. Evans and Heller, 1994] consider magnetite grains inside magnetotactic bacteria, based on the results from the western Loess Plateau.

In this study, linear correlations between magnetic susceptibility and some magnetic parameters indicate the two magnetic properties of loess and paleosol in the western Loess Plateau enable a division into two magnetic components. Only one component, which is named ‘enhanced component’, contributes to the magnetic enhancement. Since it has uniform grain-size distribution (regardless of the degree of magnetic enhancement), a strong grain-size control is required and chemically formed magnetic grains are unlikely therefore. The different origins (and different preservation condition) of the pedogenic magnetite between the western and central Loess Plateau are affected by the different climate.

The second target is a climatic record in the lacustrine sediment from Er-hai Lake in the southeastern margin of the Tibetan Plateau. In this region, a number of studies using the cores of sediments which has accumulated during last tens of thousands of years [e.g. Lin et al., 1986; Sun et al., 1986; Miura et al., 1995; Hodell et al., 1999]. Such studies have generally taken notice of the climatic change at Pleistocene-Holocene boundary. However, climatic change during Holocene has not been received much attention although the records during Holocene were available. Here, a cooling event at around 3.5 ka is pointed out by rock-magnetic and palynological analyses. It is correlated to the glacier advance on the Tibetan Plateau at the Neoglacial period.

In this part, the transition from greigite zone to magnetite zone, which is detected by rock-magnetic analyses, plays an important role in paleoclimatic implication. Thermomagnetic analyses and magnetic hysteresis measurements reveal the existence of the SP (superparamagnetic) greigite component and the SSD (stable single-domain) magnetite component. The change of the mixing ratio is observed both in the thermomagnetic and magnetic hysteresis curves in the following way. On the thermomagnetic curve, greigite has a characteristic ‘hump’ between 400–540°C. The gradual change in the sizes of the humps suggests the change of the mixing ratio. Potbellied hysteresis loops indicates the mixing of SSD grains with lower saturating field and SP grains with higher saturating field. The transformation of hysteresis loops from SP-like through potbellied to SSD-like shows the change of mixing ratio.

Two magnetic components were detected in both loess-paleosol sequence in western Loess Plateau and lacustrine sediment from Er-hai Lake. Furthermore, the origins of two components are somewhat similar in both cases: one is the detrital component and the other is the newly formed component according as the paleoenvironment. Clarification of the origins of magnetic minerals enables more detailed estimation of paleoenvironment than reliance on a single magnetic parameter such as magnetic susceptibility.

Chapter 2

Magnetic grain-size distribution of the enhanced component in the loess-paleosol sequences in the western Loess Plateau of China

2.1 Introduction

The Chinese loess-paleosol is one of the representative Quaternary sediments that have recorded the paleoclimatic changes in the continental area. Magnetic properties of the loess-paleosol sequence have been considered as one of the best paleoclimate proxies [Maher and Thompson, 1999a]. Heller & Liu [Heller and Liu, 1984; Heller and Liu, 1986] reported good correlation between the low-field magnetic susceptibility of the loess-paleosol sequence and the marine oxygen isotope record. Since their pioneer studies, low-field magnetic susceptibility profile has been regarded extensively as a convenient proxy for paleoclimate determination [Liu et al., 1992; Heller et al., 1993].

Several models to interpret magnetic enhancement of paleosol have been proposed so far. Now most researchers accept the idea of a pedogenic origin for very fine magnetite/maghemite as the source of magnetic enhancement [e.g. Heller and Evans, 1995; Maher and Thompson, 1999a].

Magnetite particles with variable grain size distribution are observed by means of electron microscopy, and regarded them as the pedogenic component of the loess-paleosol sediment from the central Loess Plateau [Maher and Thompson, 1992]. In contrast to them, we report magnetite with restricted grain size distribution as the pedogenic component of loess-paleosol sediment in the western Loess Plateau, whose present climate is more arid and colder than that of the central Loess Plateau, based on two contrasting data sets.

2.2 Samplings and laboratory methods

Loess and paleosol samples were taken from the Beiyuan section of Linxia City and the Shajinping section of Lanzhou City, Gansu Province, China (Fig. 2.1). The Beiyuan section is on the third terrace of the Daxia River (a tributary of the Yellow River) in Linxia City. This section is 38 m in height and covers the last 140 ka [Fang et al., 1994]. It consists of Holocene paleosol S0, Malan loess L1 and four embedded Sm paleosols, and S1 paleosols (Fig. 2.2(a)). The Shajinping section is on the second terrace of the Yellow River in Lanzhou City. This section is 22 m in height and covers the last 60–75 ka [Fang et al., 1999b; Yamada et al., 1999]. It consists of Holocene paleosol (S0) at the top, Malan loess (L1) and weak embedded paleosols (Sm series), and alluvial loess at the base as shown in Fig. 2.1(a).

We collected samples using small plastic boxes at both sites. After removing weathered surfaces of the outcrops as deep as 10 cm, cubic plastic boxes (7 cm³) were pressed down into the outcrop at every 5 cm intervals and then recovered. At the Shajinping section, the some samples were collected at much smaller intervals of 1.7 ~ 2.5 cm.

Low-field magnetic susceptibility (χ) was measured with a Kappabridge KLY3 susceptometer and normalized to mass of the sample. High field slope correction for paramagnetic and antiferromagnetic contribution was not adopted. Anhysteretic remanent magnetization (ARM) was acquired in a peak alternating field of 100 mT and a steady field of 0.1 mT using a 2G Enterprises model 600 degausser. ARM was measured using an ScT C-112 cryogenic magnetometer and converted into ARM susceptibility (χ_{ARM}). Isothermal remanent magnetiza-

tion (IRM) was acquired at 1.2 T using an electromagnet. Some samples were not completely saturated in the magnetic fields up to 1.8 T. As Chinese loess-paleosol contains antiferromagnetic minerals such as hematite and goethite [Eyre and Dickson, 1995], it is difficult to completely saturate samples. In this study, IRM acquired at 1.2 T is defined as “saturated” IRM (SIRM). This means that the magnetic grains with high coercivity are ignored.

Thermal demagnetization of low-temperature remanence was carried out with a Quantum Design’s Magnetic Properties Measurement System (MPMS). SIRM was acquired after “zero-field cooling” (cooling from room temperature to 10 K in 0 T), then the decay of SIRM was measured during the heating of the sample up to 300 K.

Magnetic hysteresis measurements were performed with a MicroMag alternating gradient force magnetometer. Saturation magnetization (M_s) and saturation remanence (M_{rs}) were calculated from each hysteresis loop with a maximum field of 1 T. Remanent coercivity (B_{cr}) and coercivity (B_c) were calculated from stepwise demagnetization curve of IRM acquired at 1 T.

2.3 Results

Fig. 2.2 illustrates the profile of the magnetic parameters for the Beiyuan section. χ is generally variable at the paleosols. The most part of L1 loess shows stable low value. S0 and S1 show χ four times larger compare to L1 loess. χ_{ARM} and SIRM show similar fluctuation pattern with χ . χ_{ARM} of S0 and S1 is ten times as large as that of L1. SIRMs of S0 and S1 are twice as large as that of L1.

χ , χ_{ARM} , and SIRM values of the S0, Sm, and S1 paleosols from the Beiyuan section are larger than those of the L1 loess. The idea that the susceptibility enhancement is due to the pedogenic formation of magnetic minerals [Zhou et al., 1990; Maher and Thompson, 1992] is not inconsistent with the observed difference, because these magnetic parameters are depend on the amount of the magnetic minerals.

Fig. 2.3 shows the relations between χ , χ_{ARM} , and SIRM. Excellent correlations among these parameters are shown. Similar correlation between low-field magnetic susceptibility (χ_{LF}) and the frequency dependence of the low-field magnetic susceptibility ($\Delta\chi$) is pointed out [Forster et al., 1994]. Their idea is that the mixing of two independent sources; one is the “background susceptibility”, which is expressed as the intersection of the χ_{LF} - $\Delta\chi$ plot, and the other is the “pedogenic magnetic fraction”, which is calculated as the slope of the χ_{LF} - $\Delta\chi$ plot.

The idea can be applied to the correlations of χ_{ARM} and SIRM to χ . However, the absolute values of the magnetic parameters for the background component are indeterminable. As for χ_{LF} - $\Delta\chi$ plot, the background susceptibility can be determined as the intersection of the χ_{LF} - $\Delta\chi$ plot on the assumption that the background component contains no frequency-dependent grains, i.e. the grains

of which size are about the stable single domain (SSD) / superparamagnetic (SP) threshold. However, such assumption cannot be applied to the correlations between χ_{ARM} , SIRM, and χ . In this study the lowest value of each parameter is defined as the value of background component for convenience.

The background component is constant in quantity regardless of the degree of magnetic enhancement. The magnetic parameter of the background component is $\chi = 2.7 \times 10^{-7} \text{ m}^3/\text{kg}$, $\chi_{ARM}/\chi = 1.7$, and $\text{SIRM}/\chi = 1.5 \times 10^4 \text{ A/m}$. The enhanced component has specific values of χ_{ARM}/χ and SIRM/χ , and the concentration of magnetic minerals increases with the magnetic enhancement. The χ_{ARM}/χ and SIRM/χ values of the enhanced component are 5.9 and $6.1 \times 10^3 \text{ A/m}$, respectively. S0, Sm, and S1 paleosols have identical enhanced component, as the data of three paleosols fall on a consistent regression line of Fig. 2.3. Larger χ_{ARM}/χ value and smaller SIRM/χ value in the enhanced component suggest the dominance of finer-grained magnetic grains in the enhanced component. The dominance of fine-grained magnetic grains is also implied by high M_{rs}/M_s , low B_{cr} , and B_c of high χ samples (Fig. 2.4). Presence of the pristine loess component and increase in pedogenic material in paleosols is also pointed on the basis of hysteresis curves [Fukuma and Torii, 1998].

Larger χ_{ARM}/χ value and smaller SIRM/χ value in the enhanced component suggest the dominance of finer grains of magnetic minerals in the enhanced component. Pedogenic origin of magnetic minerals which have a grain-size distribution shifted to finer size is favorable to explain the observed magnetic parameters. Further, the constant values of these ratios imply that the grain size distribution does not change with the increase in concentration.

The enhanced component is considered to be plenty of ultrafine magnetic grains [Zhou et al., 1990; Forster et al., 1994]. Superparamagnetic grains have no remanent magnetization at room temperature, but below the blocking temperature they are able to have stable remanent magnetization. When a grain is given IRM at low temperature and warmed to room temperature, it loses IRM at its blocking temperature. The blocking temperature is expressed as a function of the grain size [Néel, 1949]. Therefore the unblocking curve of low temperature IRM can suggest the grain size distribution of the magnetic minerals.

The thermal unblocking curves of low-temperature IRM (Fig. 2.5(a)) have a tendency similar to χ_{ARM} and SIRM; that is, as higher χ the sample has, as upper the curve locates in the diagram. It is also explained by the combination of two components. Namely, the remanence at each temperature step is well correlated to χ (at room temperature) and the remanence at each temperature for the two components can be calculated from the regression line. For example, the correlation between χ and the remanence at 80 K (Fig. 2.5(b)) is explained by the mixture of the background component ($5.1 \times 10^{-3} \text{ Am}^2/\text{kg}$) and the enhanced component (remanence/ $\chi = 9.7 \times 10^3 \text{ A/m}$). The unblocking curves for the two components are compilations of similar calculations at every 5 K (Fig. 2.5(c)).

For the curve of the background component, the step around 100 - 120 K is remarkable, which suggests the Verwey transition of slightly oxidized magnetite [Özdemir et al., 1993]. For the enhanced component, decrease in remanent magnetization at the Verwey transition is difficult to observe. The oxidation to $z = 0.3$ essentially suppresses the Verwey transition [Özdemir et al., 1993]. Therefore, magnetite in the enhanced component is thought to be almost fully oxidized into

maghemite.

It is notable that the unblocking curves for all of measured samples can be explained as combinations of the curves for the background and enhanced components. Especially, the curve for the enhanced component is independent on the degree of soil formation, although the amount of the enhanced component increases with the degree of soil formation. Accordingly, the grain size distribution of the enhanced component is independent on the degree of soil formation.

In the Shajinping section, χ , χ_{ARM} , and SIRM of loess-paleosol sequence are almost uniform and low (Fig. 2.6): $\chi = 2.7 \times 10^{-7} \text{ m}^3/\text{kg}$, $\chi_{ARM} = 4 \sim 5 \times 10^{-7} \text{ m}^3/\text{kg}$, and $\text{SIRM} = 4 \times 10^{-3} \text{ Am}^2/\text{kg}$. Such low values are in common with the background component of the Beiyuan section (Fig. 2.6). The existence of the common background components in the two sections and the absence of the enhanced component can explain the magnetic properties in the Shajinping section. The identical background component from the two sections within a distance of 100 km can be expected if the grains of the background component were transported to these areas by wind and not influenced by the soil formation.

2.4 Discussion

One possible process of the magnetic enhancement is chemical formation of magnetic minerals. Such example is the biologically induced mineralization [Moskowitz et al., 1993], which is chemical formation of magnetic grains under the extracellular condition of iron-reducing bacteria. It was proposed as the formation process of fine magnetite grains in topsoils from England [Dearing et al., 1996].

However, if the magnetic grains of the enhanced component were chemically produced, shift of grain-size distribution in proportion to the magnetic enhancement could occur. If we suppose that the enhanced component is formed by oxidation of some source material, the size of oxidizable grains can vary according to the condition. In another case where fine magnetic grains act as nuclei and grow coarse, the shift in grain-size may also occur. Such variable grain-size distribution is reported from the central region of Chinese Loess Plateau by means of electron microscopy [Maher and Thompson, 1992].

The topsoils from England are variable in magnetic grain-size distribution because they are variable in χ_{FD} [Dearing et al., 1996]. Changes in magnetic grain-size in the pedogenic fraction in paleosols are reported from loess-paleosol sequences in Tajikistan and East Europe [Forster and Heller, 1997; Jordanova and Petersen, 1999]. Inorganic formation of the magnetic minerals is a possible process for the magnetic enhancement at many sections.

However, the size-controlled pedogenic fractions, such as the enhanced component of the Beiyuan section, may not be produced by an inorganic formation process.

One possible origin of the size-controlled magnetite grains is magnetosomes synthesized by magnetotactic bacteria. The existence of magnetosomes has been inferred in a previous study [Evans and Heller, 1994]. The magnetosomes are formed within cells of magnetotactic bacteria and feature well-controlled grain size bounded by cell dimension. Magnetotactic bacteria are identified in sediments of various environments, including loess-paleosol sediment in the central Loess Plateau [Jia et al., 1996] and soil in Southern Bavaria [Fassbinder et al., 1990]. Thus, it is possible that magnetotactic bacteria had once existed in ancient soil and produced magnetosomes in the western Loess Plateau.

The most straightforward evidence for the origin of the enhanced component is photographs by means of an electron microscope, which can clarify the identical shape and arrangement of fine magnetic minerals. However, such photographs are only taken from well-developed paleosol samples from the central Loess Plateau. Electron microscopic photographs of paleosol in the western Loess Plateau will reveal the origin of the enhanced component in this region.

In the Shajinping section in Lanzhou City, almost no magnetically enhanced component was found in the paleosol horizons. However, eighteen weakly- to moderately-developed paleosol horizons are identified [Fang et al., 1999b]. The color reflectance data [Yamada et al., 1999] also indicates soil development. Such situation may have occurred at some levels (for example, 8–17 m and 28–35 m) of the Beiyuan section. While χ is almost constant, narrow variation of color reflectance is observed [Yamada et al., 1999]. Moreover, stable χ in loess horizons are commonly observed at several sections in the western Loess Plateau, for example, at Xining [Hunt et al., 1995], Baicaoyuan [Evans and Heller, 1994], and

Huanxian [Zheng et al., 1995].

Such observations imply that the weak soil development can occur without magnetic enhancement. The inadequate conditions for the formation of magnetic grains with controlled grain-size distribution can be estimated.

In the same way, other possible processes for magnetic enhancement of soils may require the conditions to work. The biologically induced mineralization is considered as the dominant process for magnetic enhancement in the central Loess Plateau [Maher and Thompson, 1999a], where the weather is more humid and warmer than in the western Loess Plateau. An assumption that the pedogenic condition in the western Loess Plateau is inadequate for such processes can explain the different grain-size distribution between the western and central Loess Plateau.

On the other hand, the climatic difference between the western and central regions of Chinese Loess Plateau can also affect the preservation condition of the pedogenic fine magnetite. In the Beiyuan section, the size-controlled magnetite grains could have been suffered weathering and/or pedogenesis, and oxidized to maghemite. However, the effect of weathering and/or pedogenesis might be slight, because the grain-size distribution was maintained. In case of the central Loess Plateau, strong alteration and/or dissolution may have transformed the original grain-size distribution even if magnetite with controlled grain-size distribution had once produced.

2.5 Conclusions

The confined magnetic grain-size distribution of the enhanced component at the Beiyuan section suggests that the strongly size-controlled formation of magnetic grains produced this component. It is not compatible with variable magnetic grain-size distribution reported in the central Loess Plateau. The different climatic conditions between the western and central Loess Plateau may cause the different origin and preservation condition, and different grain-size distribution of the pedogenic magnetite.

Chapter 3

A record of late Holocene cooling event from Er-hai Lake, southwestern China

3.1 Introduction

In recent years, paleoclimatic fluctuation on the Tibetan Plateau (Fig. 3.1) during late Pleistocene to Holocene has extensively been studied. Glacier regression during hypthothermal and glacier advance during Neoglacial to Little Ice Age has been explicated [Lehmkuhl and Haselein, 2000]. However, the Holocene paleoclimatic records on the Tibetan Plateau is unevenly distributed. A number of studies on climate during Holocene are from the central part and the northwestern and northeastern edges of the Tibetan Plateau [Wang and Fan, 1987; Gasse et al., 1991; Kashiwaya et al., 1995; Jelinowska et al., 1995; Rhodes et al., 1996; Avouac et al., 1996; Fontes et al., 1996; Van Campo et al., 1996; Hui et al., 1996; Gasse et al., 1996; Lehmkuhl, 1997; Thompson et al., 1997; Liu et al., 1998; Wei and Gasse, 1999; Yan et al., 1999], while previous studies in the southeastern end of the Tibetan Plateau [Liu et al., 1986; Lin et al., 1986; Sun et al., 1986; Yu et al., 1990; Miura et al., 1995; Hodell et al., 1999] pay attention to longer span rather than Holocene. In this paper, we report rock-magnetic and palynological studies on sediment cores from Er-hai Lake, Yunnan Province. A paleoclimatic record during Holocene is reconstructed based on both results and compared with the records from the adjacent area.

3.2 Sampling and chronology

Er-hai Lake is the second largest lake in the Yunnan Province, southwestern China. It is located in the western part of Yunnan province ($25^{\circ}35'-26^{\circ}N$, $100^{\circ}10'E$), and is nearby Dali and Xiaguan cities (Fig. 3.2). It is in a fault basin which runs in a north-south direction, and its dimensions are 40 km by 3–8 km. The altitude of the lake surface is 1970 m.

On the west of Er-hai Lake is a number of tributaries from mountains higher than 3000 m, while only gradual slope extends on the east and north of the lake. The only outflow river discharges from the southwestern end of the lake where Xiaguan City places. Owing to such surroundings, the major part of the lake sediment comes from the western mountains. Reflecting the substantial water input, the lake has freshwater.

A shallow depth of about 5 m separates the lake basin into a central-to-northern part and a southern part with depths of 20 m and 10 m, respectively. Piston cores ER1 and ER3 were taken from the southern and central part of the lake, respectively [Hyodo et al., 1999].

Core ER1 is 2.41 m long and mainly composed of clay but at the base silt is dominated. Core ER3 is 3.16 m long, consisting of clay above a level of 2.17 m and silt below it. The upper 1 m of both cores contains one or two shell zone(s), and lower part of the core ER3 contains some fragments of wood and leaves. Shell and wood fragments were used for AMS ^{14}C dating.

Paleomagnetic samples using 10 cm^3 plastic cubes were collected without gaps from the whole part of both cores. A total of 98 samples from core ER1 and 135 samples from core ER3 were taken. Paleomagnetic results of the cores

were already published [Hyodo et al., 1999]. 26 samples for palynological analyses were corrected from core ER3. The level of each palynological sample corresponds to the level of a paleomagnetic sample.

In this study, we adopted the depth-age model of the same cores by Hyodo et al. (1999). AMS ^{14}C dating using shell and wood fragments calibrated using dendrochronological data gives a depth-age model which has high sedimentation rate of about 2 mm/yr within the range of 1.00–1.92 m (between 3.7–3.0 ka) and low sedimentation rates of about 0.3 and 0.4 mm/yr above and below the range respectively.

3.3 Rock-magnetic analysis

3.3.1 Laboratory Procedures

Low-field magnetic susceptibility (χ), anhysteretic remanent magnetization (ARM), and Saturation isothermal remanent magnetization (SIRM) were routinely measured for all paleomagnetic samples. χ was measured with a Bartington MS2B susceptometer. ARM was imparted in a peak alternating field of 50 μ T and a steady field of 100 mT using a Natsuhara DEM-8601 alternating field (AF) demagnetizer. SIRM was imparted in a pulse field of 2.7 T using a 2G pulse magnetizer. ARM and SIRM were measured with a 2G superconducting magnetometer or a Natsuhara SMM-85 spinner magnetometer. ARM was converted into ARM susceptibility (χ_{ARM}), namely ARM intensity divided by the intensity of the steady field.

Rock-magnetic experiments described below were carried out on selected samples.

Thermomagnetic analyses were carried out with a semi-horizontal thermobalance at School of Earth Sciences, Kyoto University. In each run, a $\sim$ 100 mg of bulk sample was heated up to about 600 or 700°C and cooled to room temperature at a rate of 8°C/min in air atmosphere. Change of induced magnetization in a field of 0.8 T was monitored at intervals of 7 seconds.

Stepwise thermal demagnetization were done in air atmosphere using a Natsuhara TDS-1 furnace. In this study, low-field magnetic susceptibility after each heating were adopted.

Magnetic hysteresis measurements were performed with a MicroMag Model

2900 alternating gradient force magnetometer at School of Earth Sciences, Kyoto University. ~ 10 mg of bulk samples were prepared for the measurements. Saturation magnetization (M_s) and saturation remanent magnetization (M_{rs}) were computed from a slope-corrected hysteresis loop with a maximum field of 1 T. Remanent coercivity (B_{cr}) and coercivity (B_c) were determined from a slope-corrected stepwise DC demagnetization curve of IRM acquired at 1 T.

Low temperature measurements were performed with a Quantum Design's Magnetic Property Measurement System (MPMS-XL) at Department of Marine Geology, Geological Survey of Japan. ~ 100 mg of bulk samples were prepared. After IRM was acquired at 10 K in a 5 T DC field, change of remanent magnetization was monitored during heating to room temperature at a typical rate of 2 K/min.

3.3.2 Core ER3

High temperature measurements

Thermomagnetic analyses were carried out in order to identify the magnetic mineral(s). The thermomagnetic curves in air of the samples below 1 m at the depth have humps (increase of induced magnetization above 400–440°C and large decrease up to 520–540°C). Above 520–540°C, induced magnetization continue to decrease at lower rate (Fig. 3.3(a)). On the other hand, the samples from the upper 1 m shows monotonic decrease of magnetization in the heating branch of the thermomagnetic curves (Fig. 3.3(b)).

There are two possible explanations for the thermomagnetic behavior above

400°C: one is the decomposition of ferrimagnetic iron sulfide (ex. pyrrhotite) [Dekkers, 1990; Worm et al., 1993], and the other is the inversion of ferrimagnetic iron oxide (ex. titanomaghemite) [Özdemir and O'reilly, 1982]. If either pyrrhotite or titanomaghemite was the dominant magnetic mineral, it would begin to alter above 300–400°C. As it was, alteration of ferrimagnetic mineral around 200°C is proved by slight winding of heating branches of thermomagnetic curves and decrease in χ after stepwise thermal demagnetization (Fig. 3.4). Thermal alteration of greigite can give an account for such behavior. Decomposition of greigite to pyrrhotite starts at $\sim 200^\circ\text{C}$, some portion of greigite changes to magnetite and maghemite above $\sim 400^\circ\text{C}$ and hematite is produced when heated to 700°C in air [Hoffmann, 1992; Roberts, 1995; Dekkers et al., 2000]. We regard the humps above 400°C as a sign of greigite.

The humps enlarge downcore. The enlargement suggests the increase of the concentration ratio of greigite to the whole magnetic minerals. The humps retain their sizes below 1.5 m at depth (except for two samples around 1.9–1.8 m), gradually shrink between 1.5–1.0 m, and finally disappear at about 1.0 m. To estimate the ratio of greigite to magnetite roughly, we calculated the height of the humps by the following procedure on the following assumptions: (1) Magnetite and greigite are the only magnetic phases which contribute the thermomagnetic curve. (2) The slope of the thermomagnetic curve of magnetite does not largely change between 400 and 540°C . (3) The magnetization of greigite component is zero at both ends of the hump (i.e. 400°C and 540°C), because greigite is completely changed to pyrrhotite at 400°C (above the Curie temperature of pyrrhotite) and it is completely changed to hematite, whose magnetization can be ignored when compared

with magnetite. As we could not measure the absolute intensity, the height of hump was normalized by the decrease of magnetization between 400 and 540°C. Downcore variation of the calculated size of humps is shown in Fig. 3.5.

Magnetic hysteresis measurements

The shapes of the magnetic hysteresis loops can be classified into three types. The samples below about 1.2 m show closed hysteresis loops (Fig. 3.6 (a)), which are typical in an assemblage of MD/SP (either multidomain or superparamagnetic) grains. The hysteresis loops of the samples from 1.2–0.4 m are ‘potbellied’ ones (Fig. 3.6 (b)). The loops of samples from upper 0.4 m (Fig. 3.6 (c)) have parallel and far apart ascending and descending branches in the middle, which are characteristic of SSD (stable single domain) magnetic grains. The transition of these types does not occur all at once: around a depth of 1.2 m whole of the loops gradually open, and around a depth of 0.4 m the middles of loops are gradually reduced.

The two common types of distorted hysteresis loops (potbellied and constricted ones) are formed as mixtures of multiple fractions which shows different magnetic behavior. The potbellied hysteresis zone overlaps with the mixing zone of magnetite (which is also present above this zone) and greigite (which is present below this zone) determined from thermomagnetic curves. Therefore, the potbellied hysteresis is formed by mixing MD/SP greigite and SSD magnetite components which are represented by Fig. 3.6 (a) and (c) respectively.

When mixing SSD and SP loops for generating distorted hysteresis loops, an SP contribution with a shallow initial slope results in a potbellied hysteresis loop

while an SP component with a steep initial slope yields a constricted loop [Tauxe et al., 1996]. Different initial slope can result from different saturating field. For example, two magnetic fractions in the Chinese loess-paleosol sequence composes constricted hysteresis loops [Fukuma and Torii, 1998]. In this case, the SSD component does not completely saturate until about 0.6 T while the SP-dominated component saturates in rather low magnetic field. As for the samples from Er-hai Lake, the SSD component (Fig. 3.6 (c)) almost saturates until 0.1 T and the MD/SP component (Fig. 3.6 (a)) does not saturate below about 0.2 T. It is because greigite has somewhat harder than magnetite [Snowball, 1991; Roberts, 1995; Dekkers and Schoonen, 1996]. The difference of saturating field explains the observed potbellied (instead of constricted) hysteresis loops.

As remarked above, the shape of the hysteresis loops gradually changes. In order to clarify the position of magnetic change, we adopted the hysteresis parameters rather than graphical comparison of the hysteresis loops. Potbellied hysteresis loops are characterized by stretched middle. Such stretch will increase both M_r/M_s and H_c . In Fig. 3.5, both M_r/M_s and H_c begin to increase at 1.5 m, and reach to the maximum value at about 0.6 m. These variation proves that the change of magnetic composition may begin at around 1.5 m, though the hysteresis loops of the samples around 1.5 m does not look potbellied prominently.

Low temperature measurements

Fig. 3.7 shows the thermal unblocking curves of low-temperature IRM. The marked loss of remanent magnetization around 90–120 K is found on every sample. It suggests the existence of slightly oxidized magnetite, which loses IRM

gradually below the Verwey transition temperature (~ 120 K) [Özdemir et al., 1993]. As greigite does not show any low temperature transitions [Roberts, 1995], the existence of greigite in the lower part of the core cannot be examined in this method.

As discussed above, the principal magnetic minerals are SD magnetite and MD/SP greigite. The samples from the lower part have large greigite content, but the behavior of low-temperature IRM reflects the minor amount of SD magnetite because MD/SP grains put out no or weak remanence signals, at room temperature at least. The loss of low-temperature IRM, excluding the large loss around the Verwey transition temperature, is caused by unblocking of SP grains. The samples from the lower part of the core loses larger part of low-temperature IRM, which supports larger magnetic contribution of SP grains.

χ , χ_{ARM} , and SIRM

Vertical variations of χ , χ_{ARM} , and SIRM are shown in Fig. 3.5. Fluctuation of these three parameters have similar trends: they are low and stable below about 1.3 m, and increase between 1.3 m and 0.5 m, where is identified as the mixing zone of SSD magnetite and MD/SP greigite. The changes of these parameters also reflect the change of magnetic mineralogy accordingly.

The most important is the different intensity of magnetic signal between SSD magnetite and SP greigite. The intensity of magnetite is about five times as large as that of greigite based on saturation magnetization. Furthermore, the ratio of remanence to saturation magnetization is high for SSD grains and low for SP and MD grains. Increasing rate of χ is not as high as χ_{ARM} and SIRM partially for

that reason.

Environmental implication

We set the following three zones by summarizing the rock-magnetic results described above. In zone A (between the bottom and about 1.4 m at a depth), SP or MD greigite is the dominant magnetic fraction. In zone B (between 1.4 m to 0.5 m), both MD/SP greigite and SSD magnetite is present, and the ratio of greigite to magnetite decreases upward. In zone C (above 0.5 m), MD/SP greigite disappears and only SSD magnetite exists as a magnetic phase.

As the catchment area of Er-hai Lake is limited to the mountains on the western side, change of magnetic mineralogy due to the change of catchment area is quite unlikely. SSD magnetite fraction, which appears in zone B and is dominant in zone C, is suggested to be the original detrital magnetic component throughout the core.

Greigite is the dominant magnetic mineral in zone A and exists up to zone B. It is often reported as a authigenic iron sulfide in lake sediments [Snowball and Torii, 1999]. The formation of greigite requires anaerobic and reductive conditions. For example, it is likely to be formed in organic-rich sediments and in sediments deposited in brackish conditions. Greigite is also formed intercellularly by magnetotactic bacteria, which are generally found in marine sediments [Hesse and Stolz, 1999]. In case of non-biogenic reduction, greigite is an intermediate product in the pathway to pyrite formation [Berner, 1984] and is believed to be formed near the sediment surface reflecting the paleoenvironment immediately after the deposition.

In zone A, the dissolution of detrital magnetite and formation of greigite is an evidence of anoxic conditions at the lake bottom. In zone C, the detrital magnetite remains and authigenic greigite is absent, which means that the condition was less anoxic. In zone B decrease of the mixing ratio of greigite to magnetite indicates the transition of lake bottom environment from more anoxic to less anoxic. The dominance of coarse SSD grains in the magnetite fraction at the lower part of zone B may reflect selective dissolution of fine magnetite. Within zone A, low values of the height of humps and slightly high values of χ , χ_{ARM} , and SIRM are exceptionally observed between 1.9–1.8 m. It may indicate a temporal interruption of the anoxic condition.

The hysteresis measurements cannot determine if the greigite component is MD or SP. However, no natural greigite with MD structure is reported: inorganic greigite in natural sediments [Roberts, 1995] and greigite magnetosomes [Pósfai et al., 1998] are typically SSD or SP. Furthermore, the SD/PSD (pseudo-single domain) boundary for greigite is as large as 1 μm [Hoffmann, 1992; Roberts, 1995], which is incredibly large for authigenic mineral. Therefore, we suppose that SP greigite exists in the lower part of the core. The large decrease of remanence during the thermal demagnetization of low temperature IRM also suggests the existence of SP phase.

3.3.3 Core ER1

Typical thermomagnetic curves of samples from core ER1 are shown in Fig. 3.8. As for humps above 400°C of the heating branch, they have the same feature as core ER3: samples from lower than 1.1 m shows the thermomagnetic curves with

humps (Fig. 3.8 (a)) and samples from upper 1.1 m show thermomagnetic curves without humps (Fig. 3.8 (b)). However, behavior around the Curie temperature of magnetite is different between cores ER1 and ER3. Sharp drop of magnetization at 580°C is conspicuous in both heating and cooling branches of thermomagnetic curves of samples from core ER1. The height of humps are calculated and shown in Fig. 3.9.

The shapes of hysteresis loops are MD/SP-like at the lower part (Fig. 3.8 (c)) and SSD-like at the upper part of the core (Fig. 3.8 (e)). In the middle part of the core (around 1.0 m deep) some of the samples show potbellied hysteresis loops (Fig. 3.8 (d)). χ and SIRM slightly rise at around 1.0 m, and χ_{ARM} increases between 1.3–0.9 m (Fig. 3.9). However, higher values of χ , χ_{ARM} and SIRM in the lower part of core ER1 (about three times as large as core ER3) dim the contrast of these parameters between the lower and upper parts of core ER1.

These results suggest that detrital magnetite in the lower part of core ER1 did not completely dissolve. The thermomagnetic curves clearly show the existence of magnetite. The remnant magnetite may contribute to χ , χ_{ARM} , and SIRM values, therefore contrast of these parameters between the lower and upper part of the core is weakened. The remnant magnetite also enfeebles the mixing effect of SSD magnetite and SP greigite on hysteresis loops.

Nevertheless, the decrease in M_r/M_s and H_c and the increase in the height of humps on the thermomagnetic curves are displayed between 1.7–0.5 m. According to the correlation of two cores [Hyodo et al., 1999], this range is correlated to zone B in core ER3. Therefore, the decrease of the mixing ratio of greigite to magnetite can be confirmed from the results of core ER1.

Anoxic condition of lake bottom is brought by stratification of lake water, and shallower depth may prevent the stratification of lake water. As the coring site of ER1 was shallower than the coring site of ER3, it is likely that the coring site of ER1 was more difficult to become anoxic than the coring site of ER3. This idea can explain the remaining magnetite in the lower part of the core ER1. Based on this idea, it may safely be said that the coring site ER3 was more chemically sensitive to the climatic fluctuation.

3.4 Palynological analysis

3.4.1 Methods

Twenty-six samples from core ER3 were processed for the pollen analysis. Pollen and spore grains were extracted from three grams of each sample by the following process: treatment with 5 % KOH, removal of humic acid and colloidal grains, collection of pollen and spore grains by flotation with a specific gravity of 2.0 g/cm³ ZnCl₂, removal of mineralogic sediments by treatment with 25 % HF, After acetolysis treatment, they were embedded in glycerine jelly. Identification and counting were performed until a total of 300 grains of arboreal pollen were achieved under an optical microscope at 400–1000× magnification. The percentages of arboreal pollen taxa are calculated on the basis of total arboreal pollen counts, and the percentages of non-arboreal pollen taxa and fern spore types are calculated on the basis of total pollen and spore counts.

3.4.2 Results

Forty-one pollen types (23 arboreal and 18 non-arboreal) were identified. We only classified spores into trilete and monolete types, although somewhat detailed types were distinguished. Identified taxa are as follows (hyphenated families or genera are difficult to be morphologically discriminated):

Arboreal pollen *Pinus*, *Abies*, *Tsuga*, *Picea*, Cephalotaxaceae-Taxaceae-Cupressaceae, *Betula*, *Colylus*, *Carpinus*, *Alnus*, *Myrica*, *Juglans*-*Pterocarya*, *Celtis*-*Aphananthe*, *Ulmus*-*Zelkova*, *Quercus* subgen. *Lepi-*

dobalanus, *Quercus* subgen. *Cyclobalanopsis*, *Castanopsis-Lithocarpus*, *Mallotus*, *Acer*, *Ilex*, Elaeagnaceae, Araliaceae, Ericaceae, and Fraxinus.

Non-arboreal pollen *Typha*, Gramineae, Cyperaceae, Liliaceae, Moraceae, *Polygonum* sect. *Persicaria*, Chenopodiaceae-Amaranthaceae, Caryophyllaceae, Curciferae, *Vitis*, *Myriophyllum*, Umberiferae, *Lonicera*, *Patrinia*, *Artemisia*, Tubliflorae, and Ligliflorae.

Three local pollen assemblage zones (ER3-I, -II, and -III) were set on the basis of the following features. The percentages of major pollen taxa, the percentages of total arboreal pollen, non-arboreal pollen and fern spores, and local pollen assemblage zones are shown in Fig. 3.10.

Zone ER3-I (below 2.1 m) is characterized by coniferous pollen of *Pinus* (15.7–32.9%) and *Tsuga* (16.5–44.4%), and broad-leaved pollen of *Quercus* subgen. *Lepidobalanus* (5.9–21.5%) and *Quercus* subgen. *Cyclobalanopsis* (10.2–24.4%). *Abies*, *Carpinus*, *Alnus* and *Ulmus-Zelkova* accompany. As for non-arboreal pollen taxa, Caryophyllaceae (1.0–4.2%) is an accompaniment.

Zone ER3-II (2.1–1.4 m) is characterized by coniferous pollen of *Pinus* (70.8–77.0%) and *Tsuga* (5.1–10.9%). *Quercus* subgen. *Cyclobalanopsis* is significantly decreased to 1.3–4.4%. *Ulmus-Zelkova* also decreased to <1% through this zone and zone ER3-III. On the other hand, *Abies* increased to 5.1–10.9%. The non-arboreal Gramineae suddenly increases at the bottom of zone ER3-II and reaches to 6.1–12.7%.

Zone ER3-III (above 1.4 m) is characterized by high percentages of *Pinus* (35.0–64.3%) and appearance of *Juglans-Pterocarya* without a break (1.8–11.2%). *Tsuga* gradually decreased from 32.5% to <10%. *Abies* decreased to

less than 3% in the upper part. For non-arboreal pollen, Gramineae comprises high percentages (5.1–15.8%) and Cyperaceae appears in most of samples, although in low percentages.

3.4.3 Interpretation

The results of our pollen analysis in Er-hai Lake suggest a cooling trend of climate in zone ER3-II. At the same time, an influence of human activity is presumable.

The high percentages (72.2–89.4%) of the arboreal pollens confirm the existence of forest around the Er-hai Lake without break. The constituents of the forest are *Pinus*, *Tsuga*, *Alnus*, *Quercus*, which are similar as those of present vegetation. Above all, *Pinus* and *Tsuga* are dominant. *Pinus* increases and *Tsuga* decreases toward the upper part of the core.

The forest which is characterized by *Tsuga* and evergreen oaks (*Quercus* subgen. *Cyclobalanopsis*) is reconstructed from zone ER3-I. In zone ER3-II, *Quercus* subgen. *Cyclobalanopsis* remarkably decreases while *Pinus*, *Abies* and Gramineae increase. *Tsuga* shows high percentage subsequently to zone ER3-I. As the altitudal ranges in Yunnan of *Abies* and *Tsuga* (except *Tsuga Chinensis* var. *tchekingensis*) are as high as 2,600–4,500 m and 2,000–3,150 m respectively [Institutum Botanicum Kunmingense Academiae Sinicae, 1984], existence of *Abies* and *Tsuga* implies a transition to a relatively cooler climate. Significant decrease of *Quercus* subgen. *Cyclobalanopsis* and increase of *Pinus* and Gramineae may be a consequence of human activities. In zone ER3-III, *Pinus* further increases, and mountainy *Abies* and *Tsuga* relatively decrease therefore. *Juglans-Pterocarya* appears in the bottom of zone ER3-III and increases toward the upper part of zone

ER3-III. The appearance of *Juglans-Pterocarya* may be a mark of human activity, but it is not a definite evidence.

3.5 Discussion

Both rock-magnetic and palynological analyses reveal environmental transitions in the middle part of the cores. Rock-magnetic analyses reveals an interruption of anoxic lake bottom condition at 3.7 ka (1.9–1.8 m in core ER3) and a transitional zone from more anoxic to less anoxic condition between 3.5–1.5 ka (1.4–0.5 m). Palynological evidence of a vegetation modification began at 4–3.5 ka (2.1–1.4 m) and continued until today. These results lead to an implication that an event which affected vegetation and lake bottom condition had occurred.

Pollen analysis suggests that human activity may be considered, but human activity cannot explain the rock-magnetic results. Rock-magnetic analysis indicates that the lake bottom condition changed to unfavorable for production of greigite, which is a negative evidence for organic pollution due to human activity. Sedimentation rate temporally increases between 3.7–3.3 ka, but returns to low value after 3.3 ka. Deforestation and cultivation of the catchment area may not exert a larger influence on the lake bottom environment at least after 3.3 ka, because they may increase the sedimentation rate.

The results of pollen analysis makes another suggestion that the climate around Er-hai Lake began to cool at 4–3.5 ka. The change of lake bottom environment can be explained in this case, because warmer climate around the lake leads to the stratification and stagnation of lake water, which lead to anoxic condition at lake bottom. At the beginning of the cooling event (4–3.5 ka), the temporal interruption of anoxic lake bottom condition (around 3.7 ka) is found. The principal transition of lake bottom condition occurred during subsequent cold period. Therefore we concluded the cooling event around Er-hai Lake occurred at about

3.5 ka.

Although some authors had reported climatic records around Er-hai Lake from Late Pleistocene to Holocene [Miura et al., 1995; Lin et al., 1986], they did not report any climatic event around 4–3 ka. It is because they set the subject on the transition from Pleistocene to Holocene, which has longer time span and greater change of climate than all climatic events during Holocene.

According to a pollen analysis at the southeastern end of the Er-hai Lake [Miura et al., 1995], six local pollen assemblage zones (zones E-I–E-VI) were determined by analyzing a 7.2 m long core. High percentages of *Pinus* throughout the core is in common with our result from core ER3. Zone E-III (4.5–2.3 m at depth), which is defined as *Pinus-Quercus-Tsuga* zone, is dated between 8750 ± 70 and 5760 ± 40 ^{14}C years B.P. The core ER3 is comparable to zones E-II and E-I, which are defined as *Pinus* zone and *Pinus-Alnus* zone, respectively. They regarded sudden increase of Gramineae (containing *Oryza*) and appearance of *Fagopyrum* recorded in zone E-I as an imprint of human activity.

Paleoclimatic transitions since 17.5 ka B.P. is estimated on the basis of stratigraphical and vegetational changes at Xi-hu Lake, which lies about 10 km north of the Er-hai Lake [Lin et al., 1986]. By 10.5 ka B.P., high mountain conifers such as *Picea*, *Abies*, and Cupressaceae had disappeared and only *Tsuga* had remained in the humid spot. After 10.5 ka B.P., *Pinus* forest extends, which implies subtropical monsoon climate like today. No certain evidence of climatic changes after 10.5 ka B.P. is found. The influence of human settlement, which began at about 3 ka B.P. on the basis of historical record, is also indistinguishable in the pollen diagram.

However, the following climate signals are reported around Kunming (about 250 km southeast of Er-hai Lake). Although the original authors has not recognized them as cooling events, they can ensure that the cooling event at about 3.5 ka is not confined around Er-hai Lake, but ranges at least to Kunming. Paly-nological analysis at the northern end of Dian-chi ('i' in Fig. 3.1) near Kunming city [Sun et al., 1986] is similar to the results from Er-hai Lake: according to the pollen percentage diagram, increase in *Pinus* during last 7,000 years and sudden decrease in *Quercus* and *Cyclobalanopsis* at about 4 ka is visible. Sudden decrease in CaCO_3 concentration and increase in magnetic susceptibility around 3–2 ka are recorded in lake sediment cores from Xingyun-hu and Qili-hu ('j' in Fig. 3.1) [Hodell et al., 1999]. The decrease in CaCO_3 indicates the weakening of organic productivity, and may be correlated to the cooling event recorded in Er-hai Lake. The change of magnetic susceptibility is the same increasing trend as that in Er-hai Lake. Hodell *et al.* [Hodell et al., 1999] inferred enhanced input of eroded soils as a consequence of human activity, however.

Simultaneously with the cooling event in Yunnan Province, climatic fluctuation occurred on the Tibetan Plateau. Wang and Fan [Wang and Fan, 1987] pointed out the beginning of the so-called Neoglacial on the Tibetan Plateau at 3000 ^{14}C years B.P. They estimated decrease of temperature by 8 °C, on the basis of Holocene paleoclimatic data. Since their review, arid and/or cold events during 4–3 ka were reported from a number of sites on the Tibetan Plateau ('a'–'h' in Fig. 3.1) [Gasse et al., 1991; Kashiwaya et al., 1995; Fontes et al., 1996; Van Campo et al., 1996; Hui et al., 1996; Gasse et al., 1996; Liu et al., 1998; Wei and Gasse, 1999]). Arid events on the Tibetan Plateau are relevant to cold

conditions, because moisture is brought into the interior of the Asian continent by Asian summer monsoon, which requires a driving pressure gradient caused by heating of the Tibetan Plateau. The duration of these events are generally as long as 1000 years, because they correspond to maximum extent of glaciers. However, It is estimated that the cold climate on the Tibetan Plateau still continued after the maximum period of Neoglacial [Lehmkuhl and Haselein, 2000]. We believe that the cooling event in the Yunnan Province is an extension of the Neoglacial cooling to the east.

The influence of the Neoglacial on the Tibetan Plateau is obscure in the eastern part of China. An *et al.* [An et al., 2000] argued that climate changes in the eastern China are controlled by the weakening of the East Asian monsoon based on a compilation of the Holocene precipitation proxies. On the contrary, in Yunnan Province present precipitation is brought by the Indian monsoon [Li and Walker, 1986]. The effect of the cooling on the Tibetan Plateau is not concealed in the Yunnan Province for that reason. Therefore, the eastern limit of the cooling incident in 3.5 ka due to the Neoglacial advance on the Tibetan Plateau is presumed as the region under the Indian monsoon. If glacier advance on the southeastern Tibetan Plateau was promoted by water supply brought by the Indian monsoon, increase in precipitation may be inferred in Yunnan Province. High sedimentation rate during 3.7–3.0 ka and increase in *Abies* (which is distributed in cold and humid region in Yunnan Province [Li and Walker, 1986]) in zone ER-II may reflect increase in precipitation.

3.6 Conclusions

Rock-magnetic and palynological analyses in Er-hai Lake revealed a cooling event at about 3.5 ka. The cooling event extends at least over Yunnan Province, because it is correlated to paleoclimatic signals from other locations of Yunnan Province. It is considered as a consequence of the Neoglacial glacier advance on the Tibetan Plateau, which reached to the peak at the same time.

Chapter 4

Concluding remarks

Two magnetic components were distinguished from both sediments. The background component which has eolian origin and the enhanced component which has pedogenic origin are revealed in the loess-paleosol sediment from the western Loess Plateau. SSD magnetite component which has detrital origin and SP greigite component which has diagenetic origin are recognized in the lacustrine sediment from Er-hai Lake. In both cases the presence/absence and the quantity of the latter component are found out to represent the fluctuation of paleoclimate, because it is formed in response to the surrounding environment.

It is difficult to distinguish multiple magnetic components using a single magnetic parameter such as magnetic susceptibility, because it reflects many factors such as granulometry, composition and quantity of magnetic minerals. Multiple magnetic components are more easily defined using response curves to the change of temperature or magnetic field, which have detailed information of magnetic minerals; for example, the thermal demagnetization curves of low temperature

IRM of loess-paleosol represent the magnetic grain-size distribution; the thermomagnetic curves of sediment from Er-hai Lake indicate the alteration of a specific magnetic mineral; and the hysteresis loops of sediment from Er-hai Lake suggest grain-size and coercivity of magnetic mineral.

It is worth mentioning that a single magnetic parameter, such as magnetic susceptibility, is not enough for estimate of paleoenvironment or paleoclimate. Magnetic susceptibility of the loess-paleosol sequences increases when the pedogenic component increases, whereas magnetic susceptibility of the lacustrine sediment from Er-hai Lake decreases when alteration of magnetic mineral proceeds. The contrasting changes of magnetic susceptibility tell a risk in regarding only one parameter as a proxy.

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Figure captions

Fig. 2.1. The locations of the sampling sites. Hatched areas show the distribution and thickness of loess-paleosol sequences in Chinese Loess Plateau. (simplified after Wang et al., 1984)

Fig. 2.2. The lithology and magnetic profiles of the Beiyuan section, Linxia. (a) Lithology (white: loess, black: paleosol). (b) Low-field magnetic susceptibility. (c) ARM susceptibility. (d) SIRM.

Fig. 2.3. (a) Low-field magnetic susceptibility vs. ARM susceptibility. (b) Low-field magnetic susceptibility vs. SIRM. Two components can be inferred as sources of magnetization.

Fig. 2.4. Magnetic hysteresis parameters. Circles: M_{rs}/M_s , squares: B_{cr} , triangles: B_c ; solid symbols: results from the Beiyuan section, Linxia, open symbols: results from the Shajinping section, Lanzhou.

Fig. 2.5. (a) Thermal unblocking curves of low-temperature IRM for samples from the Beiyuan section. (b) Low-field magnetic susceptibility vs. the remanence at 80 K, i.e. the points on (a). (c) Calculated unblocking curves for the two magnetic components. The points show the two components in (b).

Fig. 2.6. The lithology and magnetic profiles of the Shajinping section, Lanzhou. (a) Lithology (white: loess; black or grey: paleosol). (b) Low-field magnetic susceptibility. (c) ARM susceptibility. (d) SIRM. Note that horizontal scales are same as Fig. 2.2.

Fig. 3.1. The location of Tibetan Plateau and Er-hai Lake. The following are the sites of climate records in previous studies: a. Sumxi-co and Bangong-co Lakes [Gasse et al., 1991; Fontes et al., 1996; Van Campo et al., 1996; Hui et al., 1996; Gasse et al., 1996; Wei and Gasse, 1999]; b. Zhacancaka Lake [Wang and Fan, 1987]; c. Zharinamu-co Lake [Wang and Fan, 1987]; d. Siling-co Lake [Kashiwaya et al., 1995; Wei and Gasse, 1999]; e. Yangbajian [Wang and Fan, 1987]; f. Noryong-co Lake [Wang and Fan, 1987]; g. Dundle ice cap [Liu et al., 1998]; h. Qing-hai Lake [Wei and Gasse, 1999]; i. Dian-chi Lake [Sun et al., 1986]; j. Xingyun-hu and Qili-hu Lakes [Hodell et al., 1999].

Fig. 3.2. The location of the sites of cores ER1 and ER3.

Fig. 3.3. Typical thermomagnetic curves of samples from core ER3.

Fig. 3.4. Change of χ after stepwise thermal demagnetization in air.

Fig. 3.5. Downcore variation of rock-magnetic parameters of core ER3. (See text for the precise definition of the size of ‘hump’ for thermomagnetic curves.)

Fig. 3.6. Typical hysteresis loops of samples from core ER3. (a) MD/SP-like, (b) potbellied, and (c) SSD-like loops.

Fig. 3.7. Typical thermal demagnetization curves of low temperature IRM.

Fig. 3.8. Typical rock-magnetic measurements of samples from core ER1. (a,b) thermomagnetic curves, (c–e) hysteresis curves.

Fig. 3.9. Downcore variation of rock-magnetic parameters of core ER1.

Fig. 3.10. Pollen diagram of core ER3. AP: arboreal pollen; NAP: non-arboreal pollen; FS: fern spores.

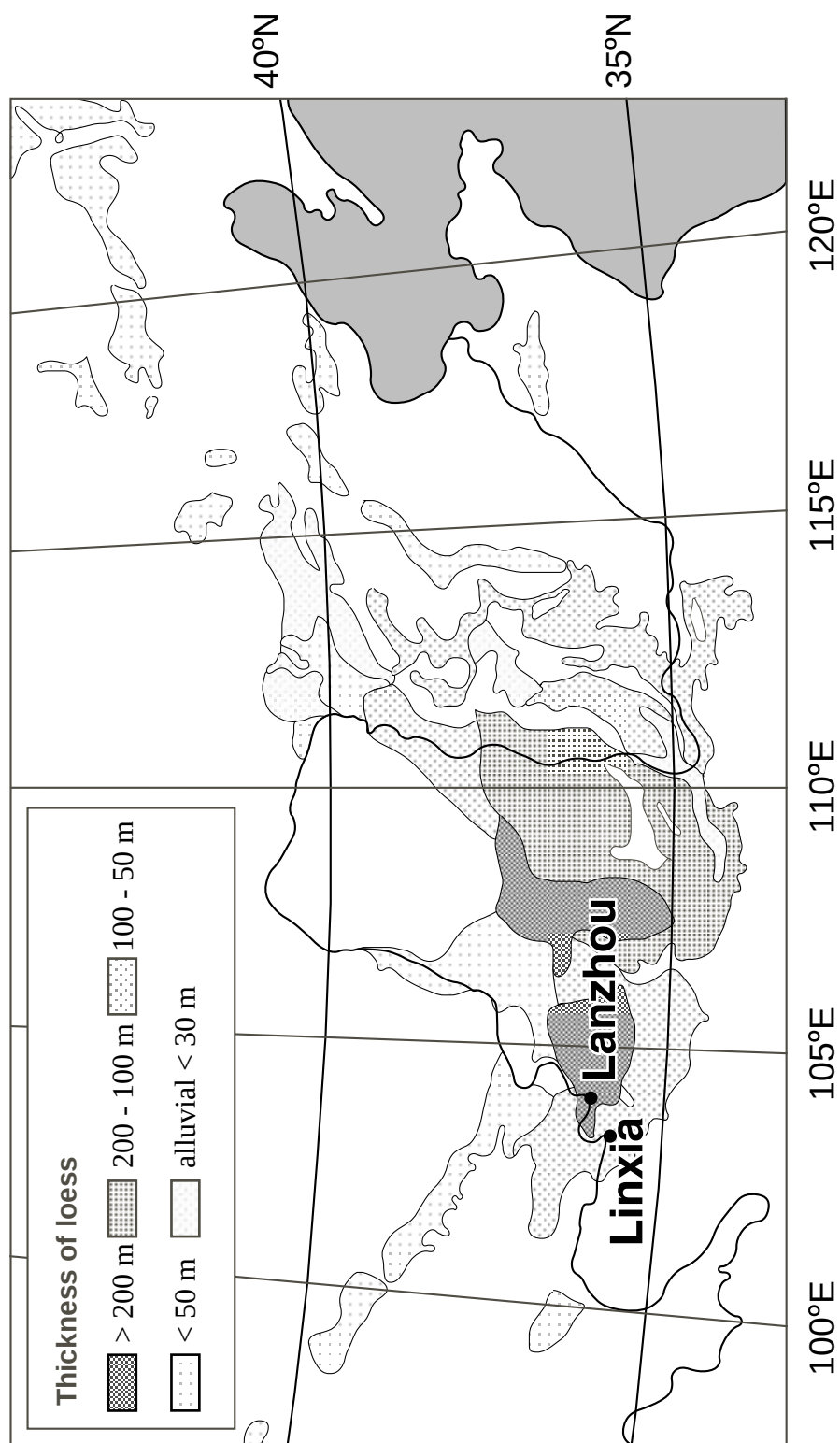


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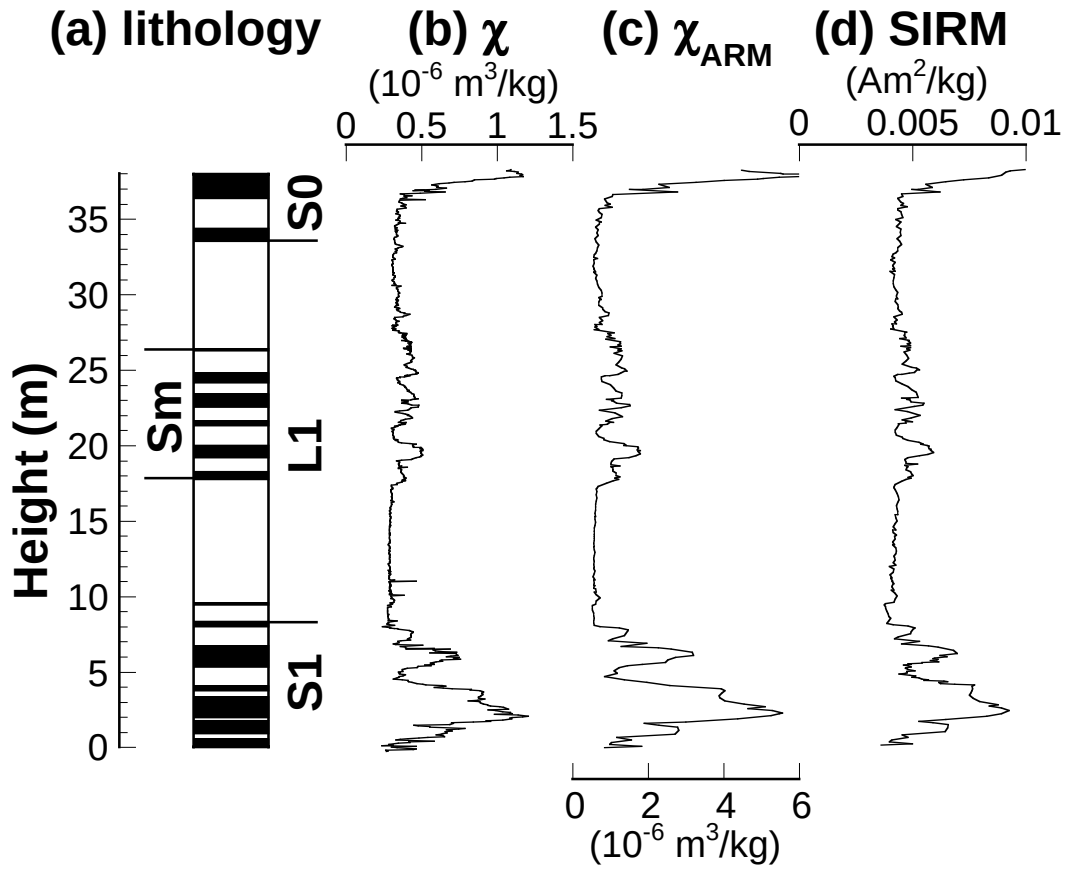


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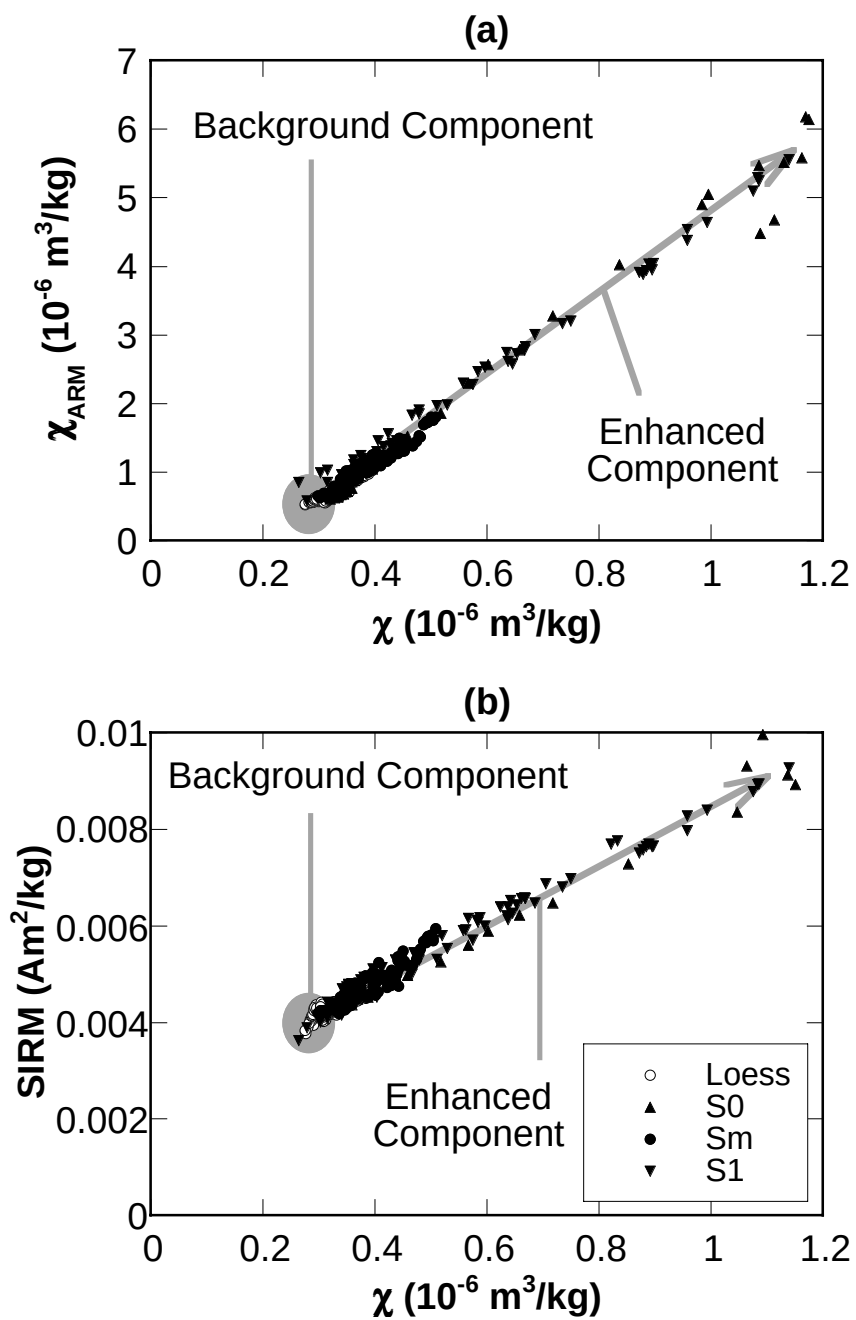


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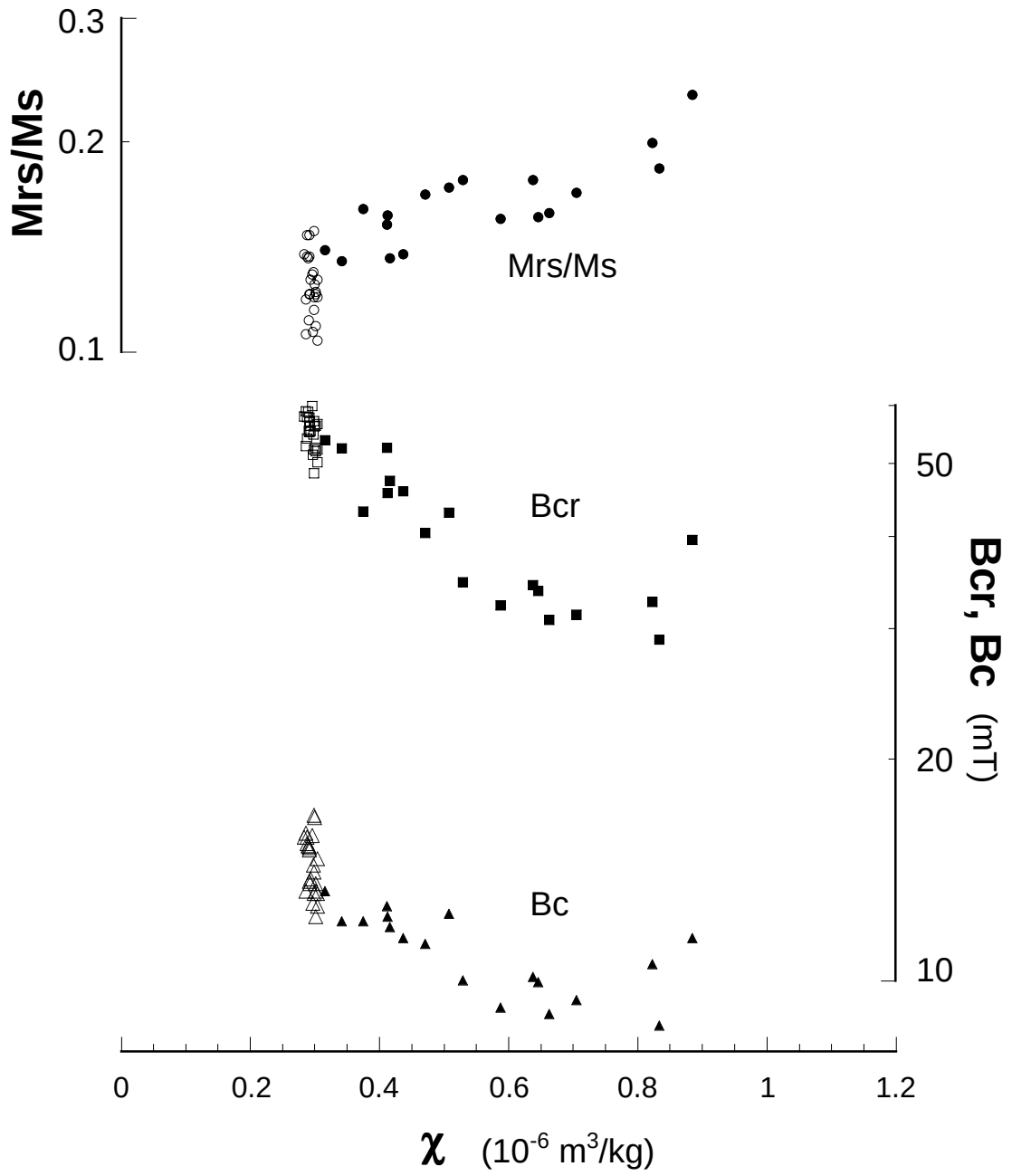


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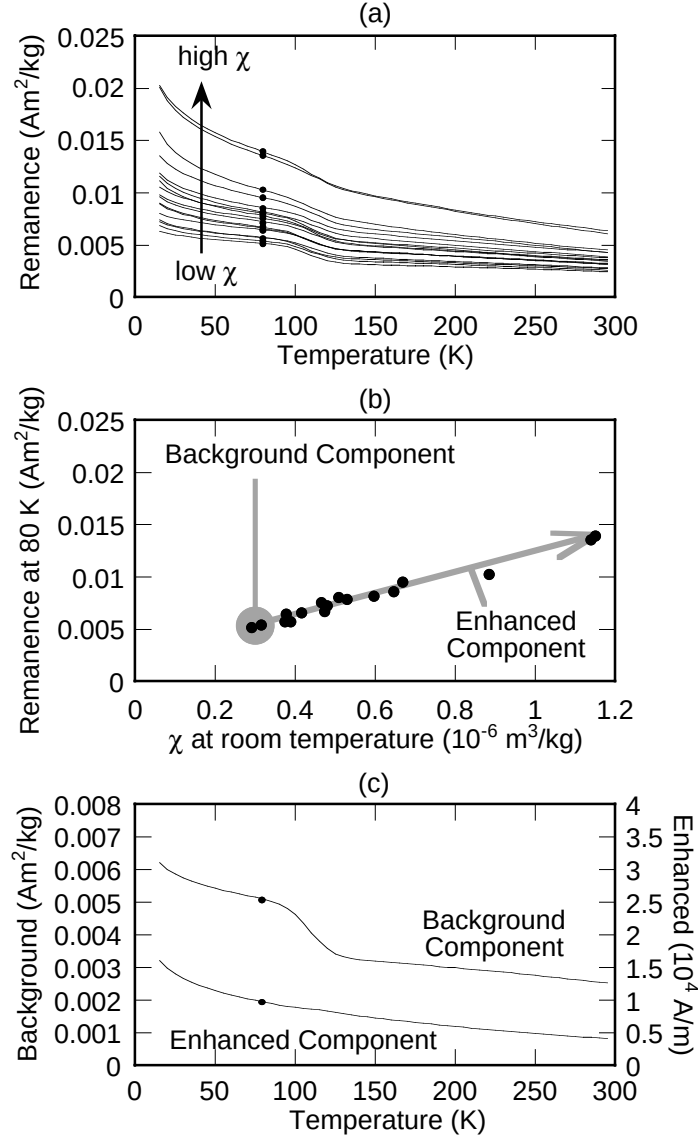


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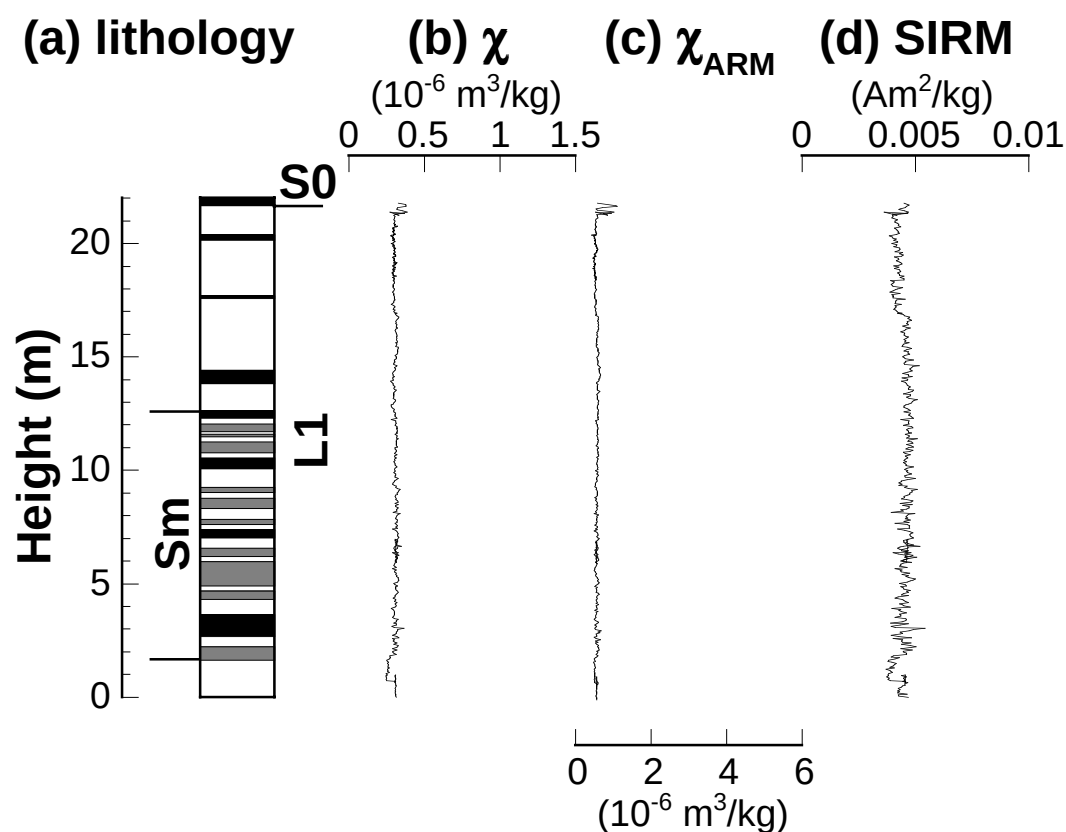


Fig. 2.6: The lithology and magnetic profiles of the Shajinping section, Lanzhou. (a) Lithology (white: loess; black or grey: paleosol). (b) Low-field magnetic susceptibility. (c) ARM susceptibility. (d) SIRM. Note that horizontal scales are same as Fig. 2.2.

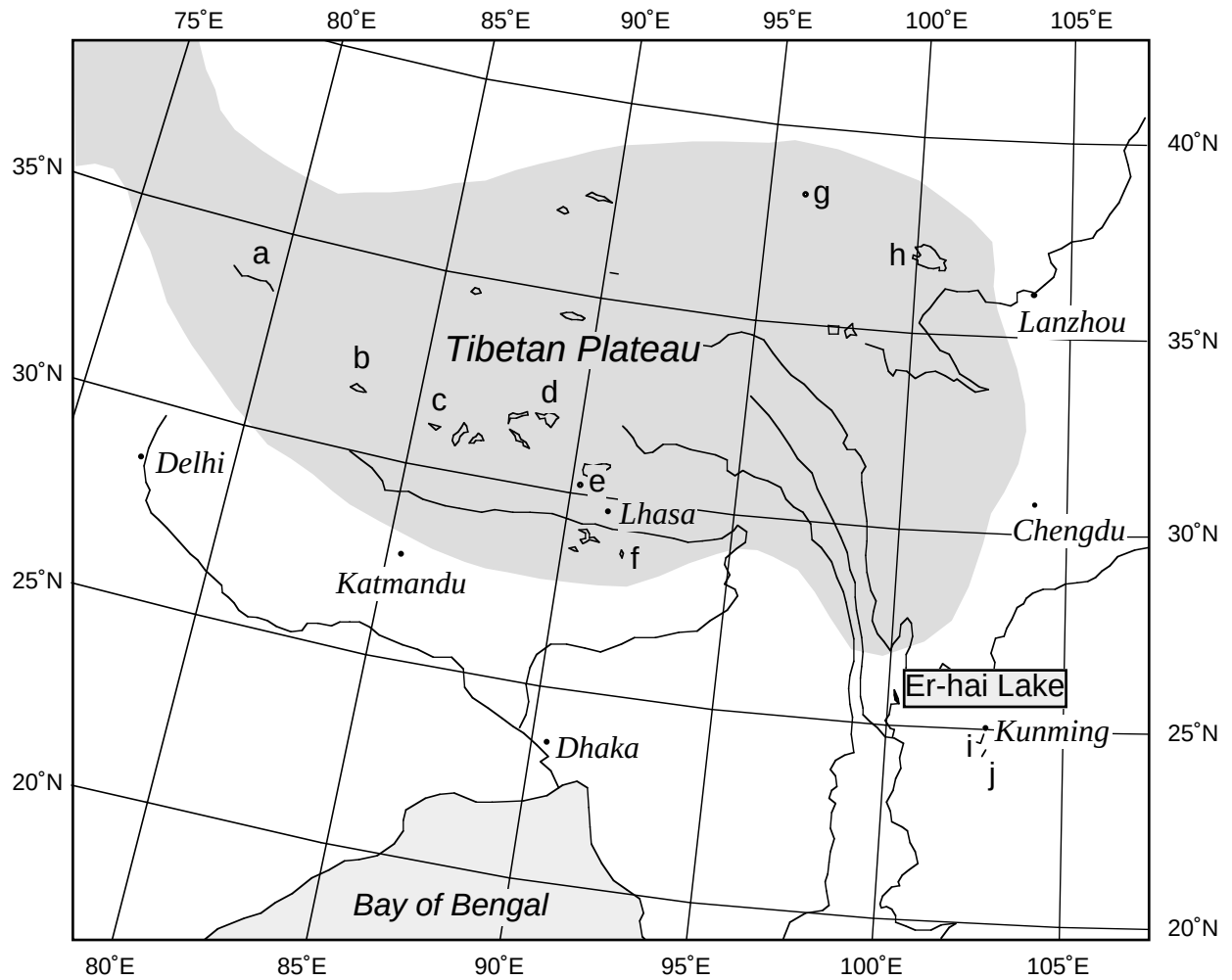


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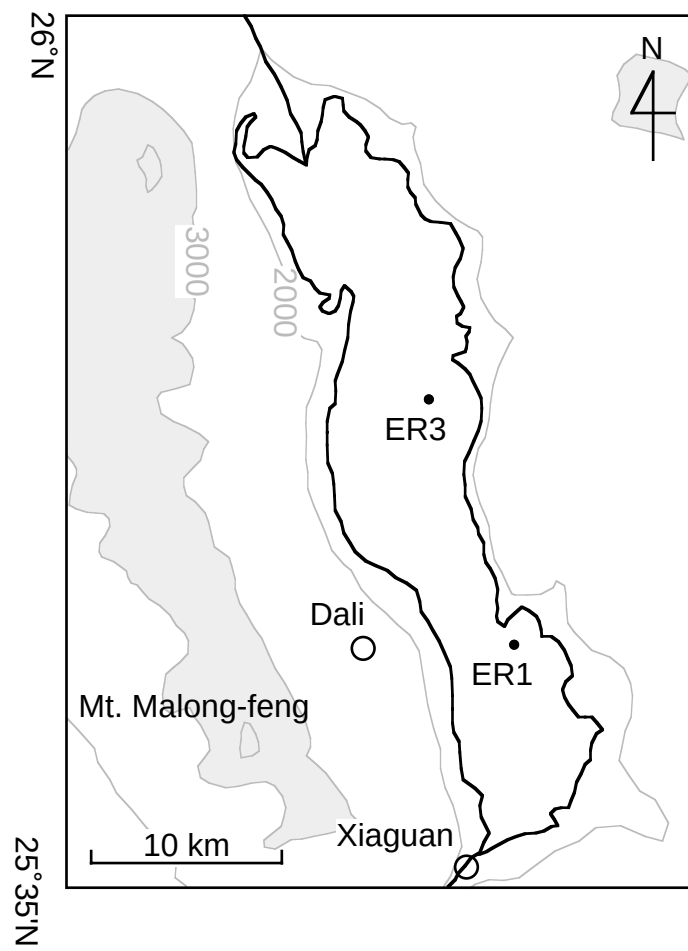


Fig. 3.2: The location of the sites of cores ER1 and ER3.

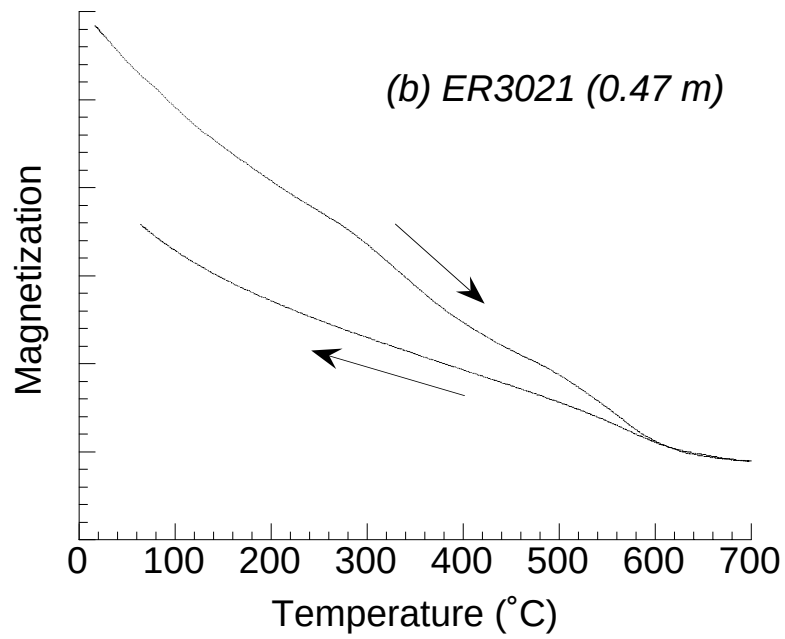
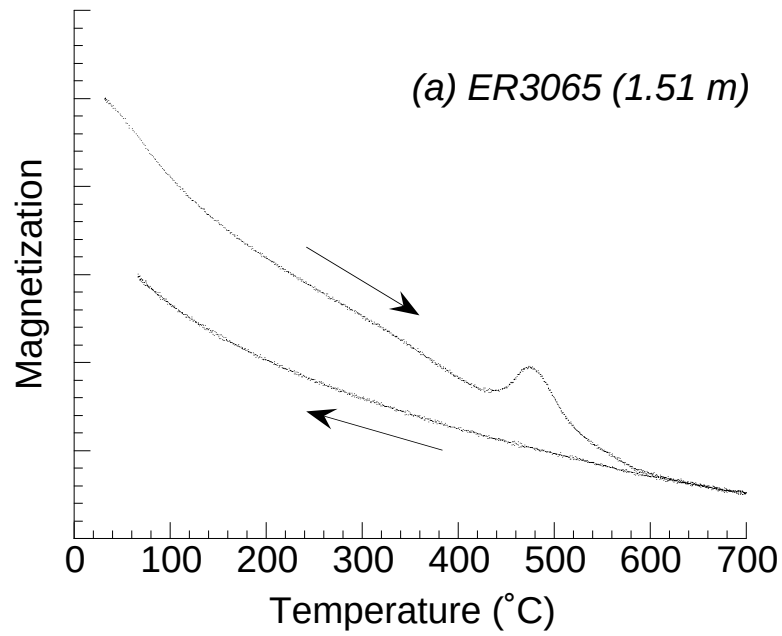


Fig. 3.3: Typical thermomagnetic curves of samples from core ER3.

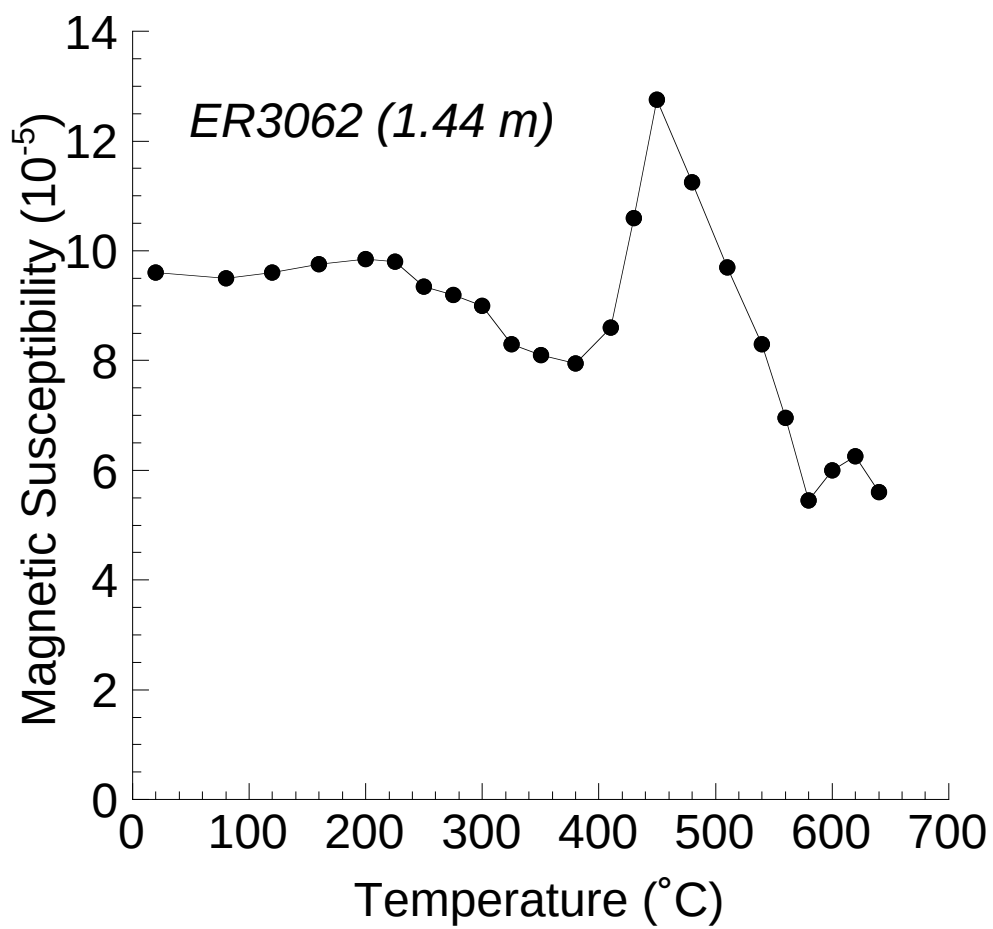


Fig. 3.4: Change of χ after stepwise thermal demagnetization in air.

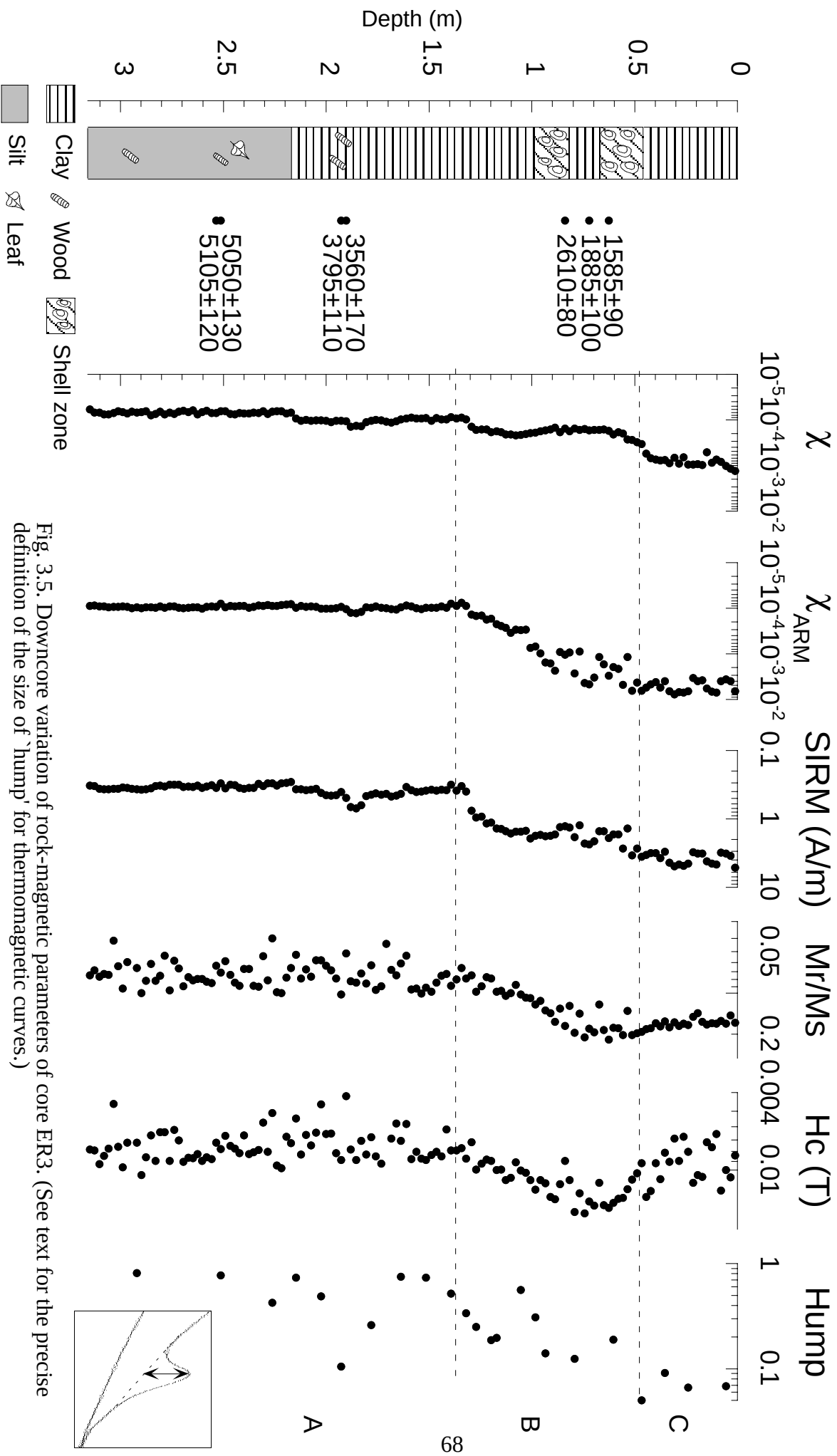


Fig. 3.5. Downcore variation of rock-magnetic parameters of core ER3. (See text for the precise definition of the size of 'hump' for thermomagnetic curves.)

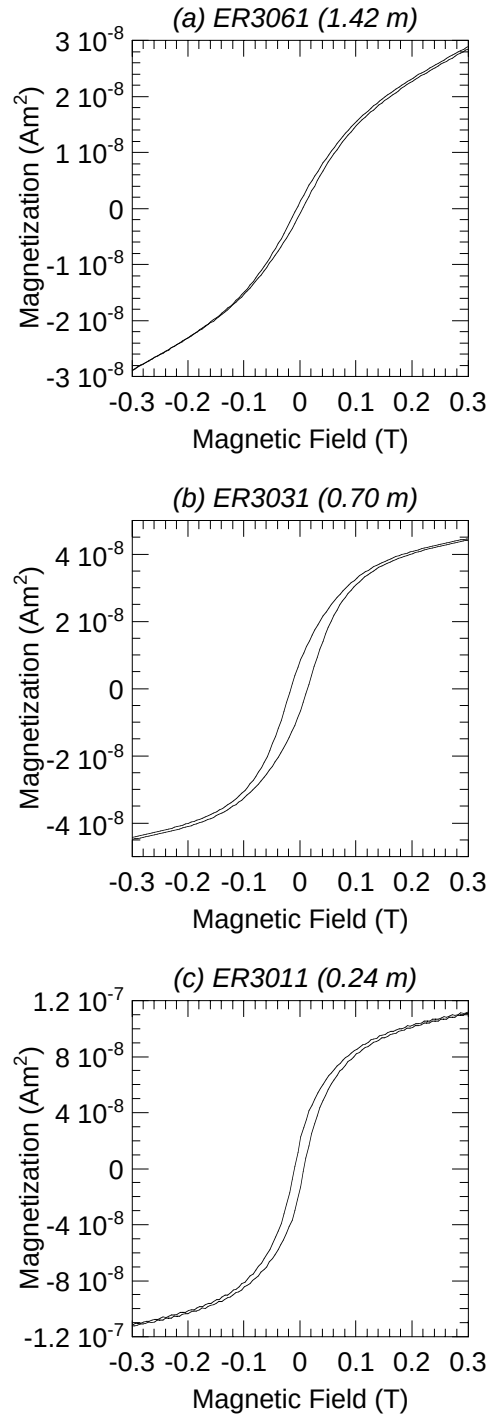


Fig. 3.6: Typical hysteresis loops of samples from core ER3. (a) MD/SP-like, (b) potbellied, and (c) SSD-like loops.

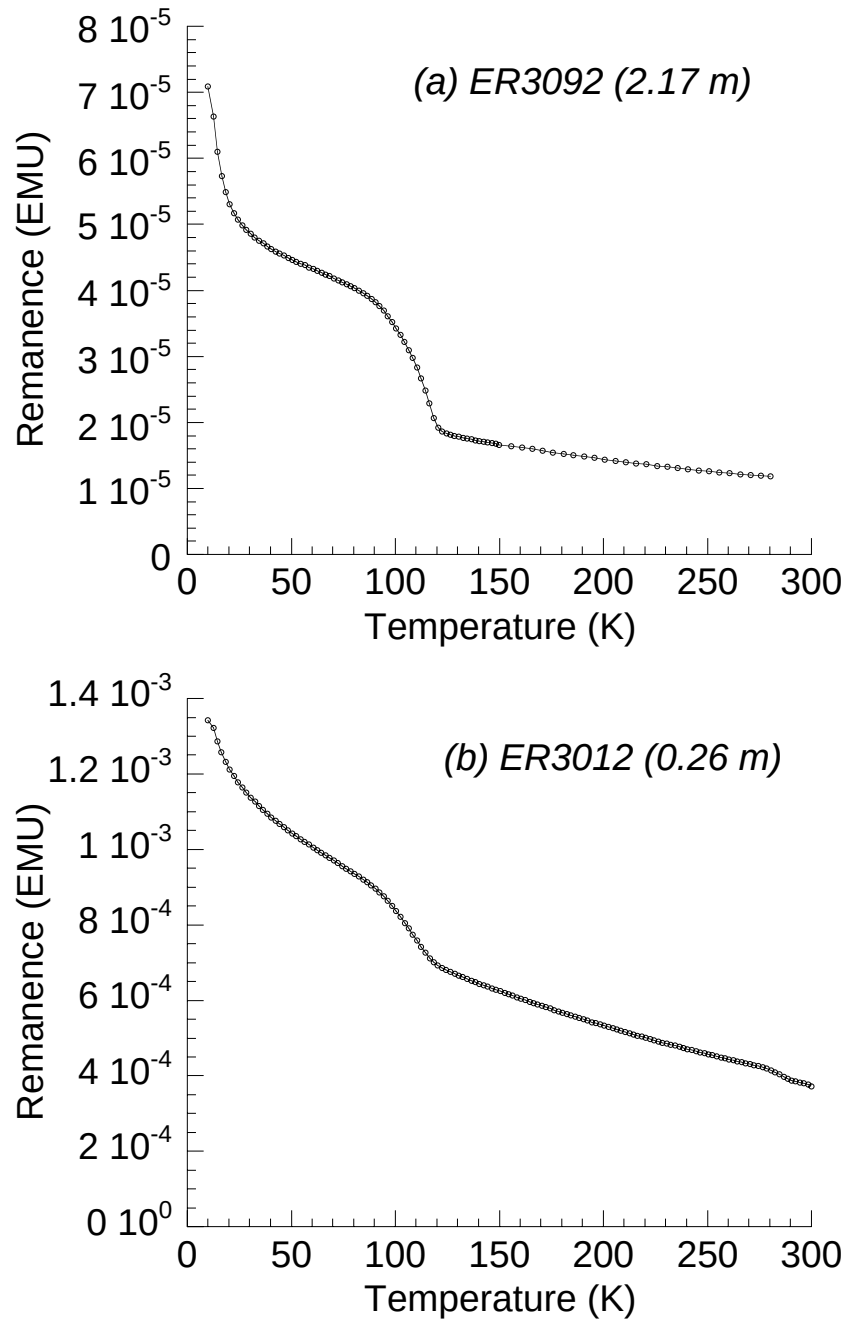


Fig. 3.7: Typical thermal demagnetization curves of low temperature IRM.

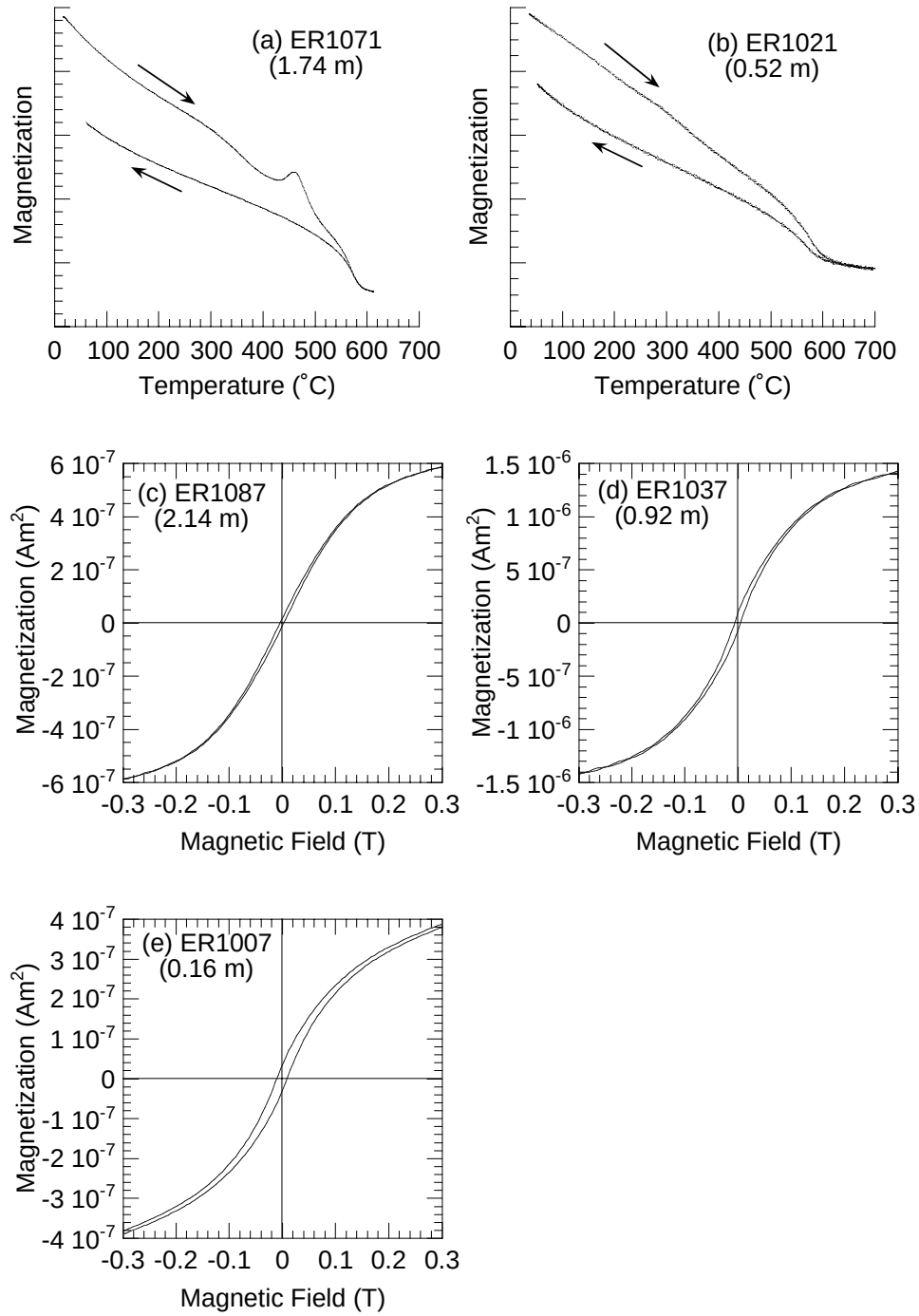


Fig. 3.8: Typical rock-magnetic measurements of samples from core ER1. (a,b) thermomagnetic curves, (c–e) hysteresis curves.

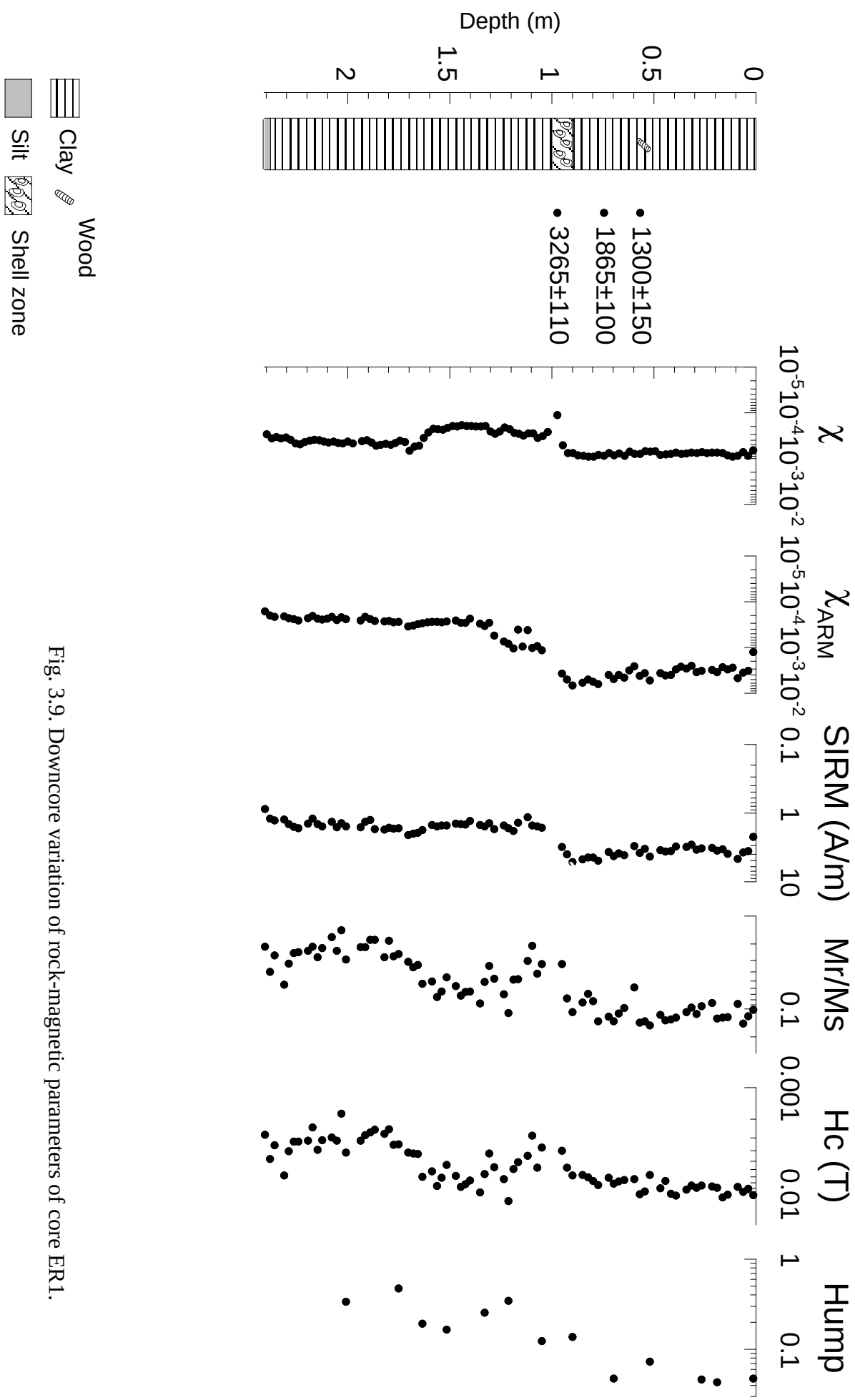


Fig. 3.9. Downcore variation of rock-magnetic parameters of core ER1.

