An Eocene–Miocene continuous rock magnetic record from the sediments in the Xining Basin, NW China: indication for Cenozoic persistent drying driven by global cooling and Tibetan Plateau uplift

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SUMMARY
Tibetan Plateau uplift and global cooling have generally been thought to have caused the drying of the Asian inland, but how and when these factors drove the aridification is unknown. The Xining Basin at the NE Tibetan Plateau received continuous Eocene–Miocene fine-grained sediments, providing an excellent opportunity to address this question. Here we present detailed rock magnetic and diffuse reflectance spectroscopy (DRS) analyses for a well-dated Cenozoic sedimentary record from the Xiejia section in the basin. Magnetic susceptibility (χ), saturation magnetization (Ms) and saturation isothermal remanent magnetization (SIRM) in this section show a long-term decreasing trend from ~52 to ~25 Ma, well coinciding with global cooling and drying in the region, and an increasing trend since ~25 Ma, which is in contrast to the further progressing aridification of the basin. Thermomorphic results and DRS-determined hematite contents suggest that the relative content of magnetite and hematite is the main control on the χ, Ms and SIRM values. We argue that the long-term Eocene–Oligocene global cooling increased the drying of the Asian inland, lowering the lake level and exposing a larger area to low temperature oxidation for longer times, thus producing more hematite and leading to the decreasing trend of magnetic concentration parameters from ~52 to ~25 Ma. An intensive uplift of the NE Tibetan Plateau since ~25 Ma, associated with a change in the sedimentary source, might be responsible for the increase of χ, Ms and SIRM after 25 Ma.

Key words: Environmental magnetism; Magnetic mineralogy and petrology; Rock and mineral magnetism; Asia.

1 INTRODUCTION
Numerical modelling and geological evidence have shown that the onset and evolution of central Asian aridification and the East Asian monsoon might have close relationships with the uplift of the Tibetan Plateau, the retreat of the Para-Tethys Sea and global cooling since the Cenozoic (Ruddiman & Kutzbach 1989; Ramstein et al. 1997; Ding & Sun 1998; Rea et al. 1998, Li & Fang 1999; An et al. 2001; Guo et al. 2002; Sun & Wang 2005; Dupont-Nivet et al. 2007; Abels et al. 2011; Hoorn et al. 2012). However, to what extent and which time interval these factors drove the aridification of central Asia is still unknown. This is largely due to a scarcity of well-dated continuous high quality Cenozoic palaeoclimatic records in Asia.

Along the northeastern margin of the Tibetan Plateau, the best exposed, most continuous and longest Cenozoic records are preserved in Qinghai Province, NW China, in which the Xining Basin (Dai et al. 2006; Dupont-Nivet et al. 2007; Fang et al. 2007) is a good representative. Detailed stratigraphic, biostratigraphic and magnetostratigraphic studies suggest that the exceptionally long continuous sedimentary succession in this basin was deposited from Eocene to the middle Miocene (Li & Qiu 1980; Qiu & Qiu 1995; Horton et al. 2004; Dai et al. 2006; Xiao et al. 2012). Previous studies indicate that the continental lacustrine stratigraphy from the Xining Basin holds a continuous sedimentary record of both Tibetan Plateau uplift (Dupont-Nivet et al. 2008; Xiao et al. 2012) and global climate changes (Dupont-Nivet et al. 2007; Xiao et al. 2010; Abels et al. 2011; Hoorn et al. 2012; Bosboom et al. 2014; Licht et al. 2014). Lithofacies and geochemical studies from the Xiejia section and the Shuiwan section demonstrated that the rapid cooling events of Oi-1 and Mi-1 and the Late Oligocene Warming were well recorded in the sedimentary archive of the Xining Basin.
lithology and facies demonstrate a slight upward trend of grain size coarsening and a long-term, upward decreasing trend of gypsum layers for Xiejia section. Below 420 m, the section mostly consists of alternating salt lake gypsum layers and salt lake marginal-distal flood plain mudflat mudstones and siltstones. Between 420 and 620 m, salt lake marginal-distal flood plain mudflat mudstones and siltstones (red beds), occasionally intercalated with thin gypsum layers (ephemeral salt lake deposition), dominate (Dai et al. 2006; Dupont-Nivet et al. 2007). Above 620 m, the sequence mostly consists of flood plain mudflat mudstones and siltstones intercalated with some channel deposits of fine sandstones with clear cross-beddings (Dai et al. 2006; Fig. 2a). A final thin gypsum layer occurs at about 500 m in the measured section (Fig. 2a). Detailed palaeomagnetic investigations from the Xiejia section indicate that the Cenozoic stratigraphy of the Xining Basin spans from about 52 to 17 Ma (Dai et al. 2006; Fig. 2b), roughly consistent with the biostratigraphic ages (Li & Qiu 1980; Qiu & Qiu 1995; Horton et al. 2004). This age determination is further confirmed by another detailed magnetostratigraphy of the Tashan section ~1 km north of the XJ section (Xiao et al. 2010, 2012; see Fig. 1b for location).

3 SAMPLING AND EXPERIMENTAL PROCEDURE

3.1 Magnetic measurements

In this study, a total of 3190 bulk samples were collected for magnetic susceptibility (χ) analyses at ca. 20 cm intervals. In addition, 14 and 230 representative samples were collected from different parts of the section for measurements of thermomagnetic curves (χ−T curves) and hysteresis loops, respectively.

Low- and high-frequency χ (mass-specific) were measured with a Bartington MS2 susceptibility meter at frequencies of 470 and 4700 Hz, respectively. The Bartington MS2 susceptometer has a sensitivity of 2 × 10⁻⁶ SI (dimensionless). From these measurements, the frequency-dependant magnetic susceptibility was calculated in percentage \( \chi_{470Hz} \), defined as \( (\chi_{470Hz} - \chi_{4700Hz})/\chi_{4700Hz} \times 100\% \) and in absolute values \( \chi_{470Hz} \) defined as \( \chi_{470Hz} - \chi_{4700Hz} \). Saturation isothermal remanent magnetization (SIRM) was imparted in a 1 T field using a magnetic measurements MMPM9 pulse magnetizer, and was measured with a Molspin Minispin magnetometer. A back-field IRM was imparted at 0.3 T (IRM₀₀₀₀₀₉₅) by reversing the orientation of the samples. The S-ratio and ‘hard’ isothermal remanent magnetization (HIRM) were determined by \(-IRM_{₀₀₀₀₀₀₉₅}/SIRM\) and \((SIRM + IRM_{₀₀₀₀₀₀₉₅})/2\) (Thompson & Oldfield 1986; Evans & Heller 2003; Bloemendal & Liu 2005), respectively. Temperature-dependant susceptibility was measured using a MFK1-FA Kapabridge equipped with a CS-4 high-temperature furnace (Agico Ltd., Brno, Czech Republic) at the Institute of Tibetan Plateau Research, Chinese Academy of Sciences, Beijing, China. High-temperature curves were recorded from room temperature to 700 °C; argon atmosphere was used to minimize oxidation. Hysteresis loops of representative samples were obtained by a Princeton Inc. alternating gradient force magnetometer (Micromag2900) at the University of Tübingen, Germany.

3.2 DRS

We collected nearly five hundred samples for DRS analysis (Barranco et al. 1989; Balsam & Deaton 1991; Deaton & Balsam 1991) from the Xiejia section at an interval of 1–2 m. In order to eliminate
Figure 1. (a) DEM and geographic map showing the location of the Xining Basin in the configuration of the Asian arid inland and the East Asian monsoon moistened region. (b) Geological map of the Xining Basin (modified from Dai et al. 2006). 1, Quaternary cover (mostly loess and fluvial deposits); 2, Linxia Group; 3, Guide Group; 4, Mahalagou Formation; 5, Honggou Formation; 6, Qijiaochuan formation; 7, undistinguished Palaeogene rocks; 8, Precambrian-Mesozoic rocks; 9, granitic rocks; 10, granodioritic rocks; 11, dioritic rocks; 12, strike-slip faults; 13, thrust faults; 14, anticlinal axes; 15, section localities.
the impact of sediment matrix, samples for DRS analysis in Xiejia section are mostly collected from the mudstones. All samples were first ground in an agate mortar. Then the fine powdered samples were pressed by hand into 20–40 mm rectangular plastic holders. Reflectance spectra of samples were measured in a Purkinje General TU1901 UV-VIS spectrophotometer with a diffuse reflectance attachment (reflectance sphere) from 360 to 850 nm at 1 nm intervals, at the Institute of Tibetan Plateau Research, Chinese Academy of Sciences. All reflectance data were analysed by first derivative (Balsam & Deaton 1991; Deaton & Balsam 1991).

4 RESULTS

4.1 Susceptibility $\chi$, $\chi_{fd}$ and $\chi_{fd}$
Values of $\chi$ in the Xiejia section range between about $1 \times 10^{-8}$ m$^3$ kg$^{-1}$ and $50 \times 10^{-8}$ m$^3$ kg$^{-1}$ (Fig. 2c). In the lower part of the section (below 420 m, corresponding to $\sim 33$ Ma), $\chi$ values are variable (the red mudstone layers have much higher $\chi$ values than the gypsum layers) and a long-term, upward decreasing trend of $\chi$ can be observed for the mudstone and sandstone layers (section I). Between 420 and 620 m ($\sim 25$ Ma), $\chi$ values are lower and generally less variable (section II). From 620 m upward (section III), values of $\chi$ show a persistent increasing trend, with a mean almost two times higher than for section II (Fig. 2c). Values of $\chi_{fd}$% are less variable (Fig. 2d). Absolute values $\chi_{fd}$ vary between 0 and $1.5 \times 10^{-8}$ m$^3$ kg$^{-1}$ (Fig. 2e) and the average value in section II ($0.3 \times 10^{-8}$ m$^3$ kg$^{-1}$) is slightly lower than in section III and in the lower part (50–200 m) of section I. In addition, we can observe higher peak $\chi_{fd}$ values in the section I than that in section II (Fig. 2e) (this means to exclude gypsum-bearing mudstone samples collected from gypsum layers).

4.2 High-temperature magnetic susceptibility ($\kappa$–$T$ curves)
According to the different characteristics of the $\kappa$–$T$ curves (Fig. 3), we find that there is a significant change in the composition of magnetic minerals across the whole section. In the lower part of the section (section I), all samples reveal a sharp loss of susceptibility near 580 °C (Figs 3a–d), indicating the presence of magnetite. In contrast, samples from the middle part of the section (section II; Figs 3e–h) the loss of susceptibility near 580 °C is less obvious, suggesting that magnetite does not provide the main control on the magnetic properties of these samples. From 620 m upwards (section III), all samples again show a sharp loss of susceptibility near 580 °C (Figs 3i–l), indicating that magnetite is the major contributor to the magnetic susceptibility.
4.3 Hysteresis parameters

The properties and shape of magnetic hysteresis loops can be used as an indicator to identify the domain state and type of magnetic minerals (Thompson & Oldfield 1986; Dunlop & Özdemir 1997; Evans & Heller 2003). After slope correction, hysteresis loops of samples from the bottom part (section I; Figs 4a–d) and the upper part (section III; Figs 4i–l) all have a steep and narrow shape and tend to close before 400 mT. Loops of samples from the middle part (section II) are relatively wide and close at higher fields near 600 mT (Figs 6e–h). The latter also show a significant wasp-waisted shape (Figs 6e–h), indicating a mixture of soft and hard minerals in these samples (Roberts et al. 1995; Deng et al. 2004).

Saturation magnetization (Ms) values are controlled by the content of ferro(i)magnetic minerals, without an influence by grain size (Thompson & Oldfield 1986; Dunlop & Özdemir 1997; Evans & Heller 2003). Fig. 2(f) shows that the Ms values for samples from section II are much lower than for sections I and III. There is a strong positive linear correlation between Ms and $\chi$ (Fig. 5a) that, in agreement with $\kappa$–$T$ curves, confirms the dominance of ferro(i)magnetic minerals. Remanent coercivities (Bcr) of samples from sections I and III are generally low, ranging at approximately 30–90 mT (Figs 2g and 4). In contrast, Bcr values of samples from section II are clearly higher, ranging from 100 to 220 mT (Figs 2g and 4). In addition, gypsum-bearing mudstone samples, collected from thick gypsum layers in section I, exhibit lower $\chi$ values and Bcr values, in sharp contrast to the negative correlation between Bcr and $\chi$ across the other part of the section (Fig. 5b). All these observations indicate a higher content of hard magnetic minerals in samples from section II.

4.4 Isothermal remanence properties

Isothermal remanent magnetization (IRM) acquisition curves for samples from sections I and III show a rapid rise below 100 mT, indicating the presence of magnetically soft components (Fig. 6). For most samples in section II only 60 per cent of SIRM is acquired below 300 mT, suggesting relatively higher coercivity phases. All samples are not fully saturated after an applied field of 1 T, reflecting the existence of hard magnetic minerals such as goethite and/or hematite (Fig. 6). SIRM is used as an estimate of the total magnetic mineral concentration with grain size larger than the superparamagnetic (SP)/single domain (SD) threshold (Thompson & Oldfield 1986; Evans & Heller 2003). The SIRM and –IRM curves for Xiejia section show quite similar variation patterns as the $\chi$ record and both can be clearly divided into three sections (Figs 7a and b). The $S$-ratio serves as a measure of the relative content of high- and low-coercivity components; samples containing relatively more low-coercivity components will have higher $S$-ratios (Bloemendal & Liu 2005). Values of the $S$-ratio for samples in Xiejia section generally span from 0.1 to 0.8 (Fig. 7c), indicating the presence of a large portion of high-coercivity magnetic minerals. The $S$-ratio values for samples from section II vary between 0.1 and 0.5 (Fig. 7c), suggesting highest concentration of high-coercivity magnetic minerals in this interval. There is a positive correlation between $S$-ratio and $\chi$ values when samples collected from thick gypsum layers are excluded (Fig. 8a). These gypsum-bearing mudstone samples, whose $\chi$ values are generally lower, show the highest $S$-ratio (0.8–0.95) and the lowest Bcr values (Fig. 5b) across the whole section, suggesting magnetically softer properties in such parts.

Figure 3. High-temperature magnetic susceptibility curves ($\kappa$–$T$ curves) of typical samples from the lower (I), middle (II) and upper (III) parts of the Xiejia section.
Eocene–Miocene continuous rock magnetic record

Figure 4. Typical magnetic hysteresis loops (after slope correction) for samples from the lower (I), middle (II) and upper (III) parts of the Xiejia section. Ms, saturation magnetization; Mrs, saturation remnant magnetization; Bc, coercivity; Bcr, remanence coercivity.

Figure 5. Bivariate plots of Ms versus $\chi$ (a), and Bcr versus $\chi$ (b) for the Xiejia section. Red solid circles represent gypsum-bearing mudstone samples collected from thick gypsum layers located at 35–50 m and 210–225 m. There is a negative correlation between Bcr and $\chi$ when gypsum-bearing mudstone samples collected from thick gypsum layers are excluded.

HIRM provides a rough estimate of the concentration of antiferromagnetic minerals (Bloemendal & Liu 2005). The HIRM record of the Xiejia section can be divided into two intervals (Fig. 7d). Below 500 m, HIRM values are higher and less variable. Above 500 m, HIRM decreases sharply and remains constant upward. These characteristics are completely different from the variations in magnetic concentration parameters $\chi$, Ms and SIRM (Figs 2c, f and 7a). These three parameters mainly reflect the content of low-coercivity ferrimagnetic minerals, but they are only weakly sensitive to hematite contents when a low-coercivity phase is present. IRM
acquisition curves of mudstone samples from Xiejia section (Fig. 6) do not saturate in a pulse field of 1 T, especially for section II where the samples show the lowest $S$-ratio values, indicating the presence of a high-coercivity phase as hematite. This conclusion is further confirmed by the results of DRS presented later.

4.5 DRS results

The DRS technique provides a useful method to determine semi-quantitatively the mass concentrations of goethite and hematite in sediments (Barranco et al. 1989; Balsam & Deaton 1991; Deaton & Balsam 1991). Previous studies have shown that hematite is characterized by a unique peak between 555 and 575 nm in the first derivative, the exact position of the peak varying with the hematite content and the degree of Al-substitution in pedogenic hematite (Deaton & Balsam 1991; Liu et al. 2011). The height of the peak increases and shifts to longer wavelength as the content of hematite increases (Deaton & Balsam 1991).

For the Xiejia section, all mudstone samples show a well-observable peak at 568 nm (Fig. 9). The lower hematite band peak position in the first derivative possibly indicates a relatively lower content of hematite in the section or an effect of Al-substitution (Balsam & Deaton 1991; Liu et al. 2011). Balsam & Deaton (1991) emphasized that the height of the first derivative peak not only decreases but also shifts to shorter wavelength as the content of hematite decreases, for example, from 595 nm for pure hematite to 565 nm for a concentration of 0.1 per cent by weight. This observation suggests that the relatively low content of hematite in the Xiejia section possibly results in a lower hematite band position in the first derivative. In section III, the band peak positions shifted subtly (by 1–2 nm) to shorter wavelengths together with a

![Figure 7](http://example.com/figure.png)

**Figure 7.** SIRM (a), $\text{IRM}_{-300\text{mT}}$ (b), $S$-ratio (c), HIRM (d) and DRS-determined hematite contents (e) of the Xiejia section and their comparison with the oxygen isotope record of deep-sea sediments (f) (Zachos et al. 2001). FDV$_{568\text{nm}}$ axis is reversed. Note the roughly synchronous variations of SIRM, $\text{IRM}_{-300\text{mT}}$, $S$-ratio and hematite contents with the oxygen isotope record for the time period between $\sim$52 and $\sim$25 Ma, and the large fluctuations of rock magnetic parameters in section I (below 420 m) that are due to the occurrence of alternating salt lake gypsum layers and salt lake marginal-distal flood plain mudflat mudstones and siltstones.
Eocene–Miocene continuous rock magnetic record

Figure 8. Bivariate plots of S-ratio versus $\chi$ (a), and Bcr versus hematite content (b) for the Xiejia section. Red solid circles represent gypsum-bearing mudstone samples collected from thick gypsum layers located at 35–50 m and 210–225 m. There is a positive correlation between S-ratio and $\chi$ when gypsum-bearing mudstone samples collected from thick gypsum layers are excluded.

Figure 9. First derivative curves for typical mudstone samples from the lower (I), middle (II) and upper (III) parts of the Xiejia section. The green line indicates the first derivative curve for 100 per cent hematite (Balsam & Deaton 1991). Note that samples from the middle (II) part of the Xiejia section have the highest first derivative values at 568 nm.

Section contain hematite and the first derivative value at 568 nm (FDV$_{568\text{nm}}$) can be therefore used as a proxy for relative changes in the mass concentration of hematite.

Figs 7(e) and 9 show that the FDV$_{568\text{nm}}$ peak is much higher for most samples from section II in comparison to samples from sections I and III. Although it varies with gypsum layers and red mudstone beds alternations, the height of the FDV$_{568\text{nm}}$ peak reveals a long-term, upward increasing trend below 420 m and a long-term, upward decreasing trend above 620 m. These observations are consistent with the rock magnetic results, demonstrating that the long-term, upward decreasing trend of $\chi$, Ms, SIRM and S-ratio from 20 m ($\sim$ 52 Ma) to 620 m ($\sim$ 25 Ma) and the long-term, upward increasing trend of $\chi$, Ms, SIRM and S-ratio from 620 m ($\sim$ 25 Ma) to 819 m ($\sim$ 17 Ma) are due to increasing and decreasing portions of hematite, respectively.

In addition, previous studies have demonstrated that goethite has a primary peak at 535 nm and a secondary peak around 435 nm in the first derivative (Deaton & Balsam 1991). However, these peaks might be obscured by the presence of hematite or possibly by clay minerals like chlorite and illite because they have characteristic peaks at similar wavelengths (Balsam & Damuth 2000). For the Xiejia section, there is no obvious peak at 535 nm (Fig. 9). Although the peaks at 440 nm can be observed, other minerals, like illite and chlorite, also can exhibit a peak around 440 nm (Ji et al. 2006). Therefore, single FDV curves cannot be used to identify goethite effectively in the Xiejia section. In addition, the $\kappa$–T curves do not show an obvious goethite signal, thus we believe that the contents of goethite do not play an important role for the observed magnetic concentration parameters variations in the Xiejia section.

5 DISCUSSION

Our rock magnetic and DRS results suggest that variation in the relative contents of magnetite and hematite provide the main control on the magnetic concentration parameters $\chi$, Ms and SIRM in the Xiejia section. There are several possible factors that might be responsible for the variation of magnetic minerals in the Xiejia section, including (1) pedogenesis in the drainage areas, (2) post-depositional dissolution of detrital magnetic minerals,
(3) low-temperature oxidation (LTO) occurring in the shallow saline lake or in the catchment area and (4) changes in provenance.

In the Xiejia section, $\chi_{d}a\%$ values generally vary between 0 and 6 per cent with a mean value of 2.6 per cent. Below 620 m, a long-term, upward decreasing trend of $\chi_{d}$ values can be observed for the mudstone and sandstone layers when excluding gypsum-bearing mudstone samples collected from gypsum layers. The decreasing trend of $\chi_{d}$ values shows a similar variation pattern as the $\chi$ record and might indicate a long-term, stepwise drying of the Xining Basin from the Eocene to the Miocene. This hypothesis is further confirmed by the long-term increasing trend of the deep-sea $\delta^{18}O$ record (Fig. 2h). All these observations suggest that superparamagnetic (SP) grains are present in the Xiejia section and their contributions to $\chi$ cannot be neglected. However, the maximal difference of $\chi_{d}$ values between section II and sections I and III is only about $0.7 \times 10^{-6}$ m$^{-3}$ kg$^{-1}$ and thus less than the 1/6 of the maximal difference of $\chi$ (25–30 $\times 10^{-3}$ m$^{-3}$ kg$^{-1}$). Therefore, we argue that pedogenically produced ultrafine ferrimagnetic minerals do not play a dominant role for the observed $\chi$ variations in our studied sequence. This is not surprising because of the arid climate in the Xining Basin during the recorded period, which is confirmed by the well-developed gypsum layers in the lower part of the section. In addition, the presence of a large portion of high-coercivity magnetic minerals, that is hematite, as suggested by the $S$-ratio ranging from 0.1 to 0.5 and the $Bcr$ values ranging from 100 to 220 mT in section II, cannot be explained by pedogenesis either. Further rock magnetic work such as decomposition of SIRM curves (Deng et al. 2006) and CBQ analyses (Verosub et al. 1995; Fine et al. 2005) might provide additional information to address the contribution of pedogenesis to $\chi$ in Xiejia section.

Post-depositional dissolution of detrital magnetic minerals during and after burial is common in lake sediments (Snowball 1993; Demory et al. 2005; Ortega et al. 2006; Ao et al. 2010). Generally, dissolution rates and effects on magnetic minerals are modulated by the availability and reactivity of both organic matters and reductants on the one hand and the competitive efficiency of microbial populations on the other (Snowball 1993). The organic carbon content throughout the Xiejia section is quite low (<0.1 per cent; Xu et al. 2009), the palaeolake was predominantly shallow (playa saline lakes; Dai et al. 2006) and sediments in other intervals are mostly highly oxidized red beds of mudflat surrounding the playa lakes and/or floodplain overbank-distal mudstones and siltstones. Therefore, we argue that the magnetic minerals in the sediments from the Xiejia section were not significantly affected by post-depositional dissolution.

We note that the deep-sea $\delta^{18}O$ record shows a long-term increasing trend from 52 to 33 Ma, and then remains less variable between 33 and 25 Ma (Fig. 2h; Zachos et al. 2001). This trend is roughly synchronous with the magnetic concentration parameters $\chi$, $Ms$ and SIRM variations in the Xining Basin (Figs 2c, f and 7a). Previous geological studies have demonstrated that the global cooling can exert a significant influence on the climatic deterioration of central Asia through the following processes: increase of the temperature gradient between the poles and the tropics (Ding et al. 2005); sea level lowering and exposure of vast continental shelves in the Pacific marginal seas (Wang 1999); weakening of oceanic water evaporation and decreasing moisture cycles in Asia (Wang 1999; Ding et al. 2005). These processes greatly enhance the continentality and result in less moisture being brought into central Asia; they finally might have caused the widespread aridity in the Xining Basin. Because LTO in the shallow lake or in the catchment area has a close relationship with the aridity (Fig. 10), we argue that increased Asian inland aridity due to global cooling since the Eocene possibly had a direct impact for the variation of ferro(i)magnetic minerals in the Xiejia section.

LTO is a process of weathering during which magnetite is altered to maghemite; at advanced stages of oxidation hematite can be formed (Sidhu 1988) which significantly lowers $\chi$ values. Oxidation is decreasing from the surface of particles towards their crystal interior, thus larger particles are less affected (Cui et al. 1994; Van Velzen & Dekkers 1999). However, it can be expected that the degree of LTO in our sediments is also increasing with the time of exposure to oxidative conditions.

During the Eocene, global warming, together with the existence of the vast Paratethys Sea in central Asia (Dercourt et al. 1993; Bosboom et al. 2011), might have intensified oceanic water evaporation and moisture cycles in Asia, causing a relatively humid climate in the Xining Basin. The lake was relatively larger and the transport distance in water increased (Fig. 10a). Consequently, clastic materials derived from the catchment area were subjected to LTO under atmospheric conditions for only relatively short times. Moreover, the relatively deep water and the cyclic gypsum beds deposited in the lake, which were chiefly determined by the water supply of solutes in continental playa-type systems, likely prevented the oxidation of the magnetic minerals in the sediments. These conditions reduced the production of hematite and resulted in higher contents of magnetite in the sediments. Gypsum-bearing mudstone samples collected from thick gypsum layers in section I, whose $\chi$, $Ms$ and SIRM values are generally lower, show the highest $S$-ratio (0.8–0.95) (Fig. 7c) and the lowest $Bcr$ values (Fig. 5b), providing further support for the LTO hypothesis.

In the Oligocene (33–25 Ma), the strong global cooling might have caused a significantly dryer climate in the Xining Basin. Consequently, the energy and potential for water transport was low and the lake area decreased significantly. Clastic materials were transported subaerially over a wider distance and were subjected to LTO for longer time. Moreover, gypsum beds disappeared and the lake was very shallow and LTO in the sediments probably became also strong. These conditions likely resulted in higher contents of hematite in the sediments (Fig. 10b).

Many lines of evidence confirm a long-term, stepwise drying of the Xining Basin from the Eocene to the Miocene. The n-alkane and palynological records from the Xiejia and Shuiwan sections in the Xining Basin revealed that the palaeoclimate in the Xining Basin experienced a long-term drying trend from 50.2 to 28.2 Ma (Long et al. 2011; Hoorn et al. 2012). Sediment lithology and facies provide further evidence (Dai et al. 2006). The slightly upward-coarsening trend of grain size of Xiejia section (Fig. 2a) reveals a progressive shrinkage and disappearance of the salt lake system and an appearance of a flood plain system; it suggests a long-term, stepwise drying of the Xining Basin from the Eocene to the early Miocene. In addition, the inhibition of gypsum accumulation in the upper part of Xiejia section, which is chiefly determined by the water supply of solutes in continental playa-type systems, also indicates a pronounced aridification of Xining Basin in the early Miocene (Dupont-Nivet et al. 2007).

The above lines of evidence indicate a further intensification of drying of the Xining Basin after 25 Ma (Dai et al. 2006; Long et al. 2011). According to our model (Figs 10a and b), this climate change should have resulted in more intensive (long-term) LTO of magnetite in the catchment area, producing higher contents of hematite in the sediment material. However, our rock magnetic
and DRS results reveal a gradually decreasing trend in the relative content of hematite from 25 to 17 Ma (Fig. 7e). This controversial trend suggests that drying in the Xining Basin was not the only mechanism for the variation in magnetic minerals in the Xiejia section.

Therefore, we argue that a change of the sedimentary source caused by tectonic uplift of the NE Tibetan Plateau might be responsible for the magnetic properties observed after \( \sim 25 \) Ma. Across the whole NE Tibetan Plateau, an angular- or pseudo-unconformity was widely observed between the Palaeogene and Neogene sedimentary rocks more than two decades ago (NBGMR1989) and its age was later confirmed more precisely by detailed palaeomagnetic and palaeosedimentary studies demonstrated that the Xining Basin was a distal sedimentary site of a large joint Gonghe-Guide-Xining Basin at the foot of the Anyemaqen Shan and East Kunlun Shan with palaeocurrents flowing northwards from the Guide Basin to the Xining Basin. From about the late Oligocene and early Miocene, probably in response to successive south-to-north synchronous growth of the Tibetan Plateau (Yin et al. 2002; Fang et al. 2003, 2012) or the rise of the Himalayas and the southern Tibetan Plateau at this time (e.g. Yin et al. 1994; Hodges et al. 1996; DeCelles et al. 2011). More locally, detailed palaeomagnetic and sedimentologic studies demonstrated that the Xining Basin was a distal sedimentary site of a large joint Gonghe-Guide-Xining Basin at the foot of the Anyemaqen Shan and East Kunlun Shan with palaeocurrents flowing northwards from the Guide Basin to the Xining Basin. From about the late Oligocene and early Miocene the Laji Shan began to partially uplift, changing the palaeocurrent directions in the Guide Basin to the south of the Laji Shan from previously northward to southwestward and later to completely southward at \( \sim 8 \) Ma (Fang et al. 2005, 2007). This uplift is further confirmed by some detrital zircon U/Pb ages from the Laji Shan (Lease et al. 2011) and more widely from apatite fission track-revealed cooling histories of rocks over the NE Tibetan Plateau (e.g. George et al. 2001; Jolivet et al. 2001; Wang et al. 2011). Thus, before the Laji Shan uplift, the clastic sediments of the Xining Basin came from the Anyemaqen Shan, which contains Triassic flysch (Clark et al. 2010). After the uplift of the Laji Shan, the sediment source for the Xining Basin changed to the more proximal volcanic rocks in the Laji Shan that contain more magnetite. In addition, mountain erosion produced a relatively larger influx of weakly weathered clastic materials due to quicker deposition. Both of these processes would decrease oxidation of incoming detrital magnetite into the basin and result in a higher content of magnetite in the sediments (Fig. 10c). Interestingly, the significantly increases of \( \chi \) after \( \sim 25 \) Ma are also investigated in the Tashan section, \( \sim 1 \) km north of the Xiejia section, and Xiao et al. (2012) argued that changes in provenance due to uplift of the Laji Shan might play a dominate role in accounting for \( \chi \) variations of the Tashan section (Xiao et al. 2012). Thus, we conclude that an intensive uplift of the NE Tibetan Plateau occurred in late Oligocene to early Miocene times which might have played a dominant role for the increase of \( \chi \), Ms and SIRM after 25 Ma.

In addition, we note that the Eocene–Oligocene transition (Dupont-Nivet et al. 2007; Xiao et al. 2010; Abels et al. 2011) was well recorded in the rock magnetic and DRS results of the Xiejia section. However, the steps of at 40 Ma, 36.7 and 34.9 Ma (Dupont-Nivet et al. 2008; Hoorn et al. 2012; Licht et al. 2014) are not significant. These characteristics possibly indicate that our rock magnetic data are not sensitive to reveal the step-wise nature in high detail. In contrast, they provide excellent opportunities to study the long-term trend of climate changes or the tectonic uplift in the NE Tibetan Plateau.

**Figure 10.** Schematic illustration of the possible driving mechanisms for explaining the \( \chi \) variation in the Xining Basin.

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**6 CONCLUSIONS**

The Xiejia section in the Xining Basin provides a first Eocene to Miocene continuous rock magnetic record for the Asian interior. It shows a long-term decreasing trend of the magnetic concentration parameters \( \chi \), Ms and SIRM from 52 to 33 Ma, followed by generally less variable and lower values between 33 and 25 Ma, and then a long term increasing trend after 25 Ma. The variation of \( \chi \), Ms and SIRM is closely related to changes of the relative contents of magnetite and hematite, and it is well matching with the global cooling record for the time period between 52 and 25 Ma.
We propose that from 52 to 25 Ma, LTO is the major mechanism for the variation of $\chi$, Ms and SIRM in the sediments of the Xining Basin, controlled by long-term, stepwise drying and driven by global cooling. In contrast, after 25 Ma, a change of the sedimentary provenance due to uplift of the NE Tibetan Plateau is a likely candidate to explain the long term increase of the magnetite-to-hematite ratio (decrease of LTO) in this interval, which does not match with the further progressing aridification.

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