Magnetostratigraphic and paleoenvironmental records for a Late Cenozoic sedimentary sequence drilled from Lop Nor in the eastern Tarim Basin

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A B S T R A C T

The Tarim Basin, one of the largest inland basins in the world, is situated in the northwestern China and to the north of the Tibetan Plateau. Continuous Cenozoic deposits have accumulated in this basin, which are crucial for investigating the growth of the Tibetan Plateau and the paleoclimatic evolution in Asian interior. Here we report the magnetostratigraphic and paleoenvironmental records for a Late Cenozoic sedimentary sequence drilled from Lop Nor in the eastern Tarim Basin. Magnetostratigraphic results show that this core has recorded a magnetic polarity sequence from C3Bn to C1n, covering an age range from ca. 7 Ma to the present. Decreased magnetic susceptibility occurred after ca. 5.6–5.1 Ma, which was interpreted to indicate an enhancement of aridity in the Tarim Basin since this period. We attribute this aridification to the combined effect of global climate cooling and the uplift of the Northern Tibetan Plateau since the late Miocene.

1. Introduction

At present, the Asian interior is characterized by a large area of arid environment, which has significant influences on the societal and economic activity in this vast area. A well constrained aridification history of this Asian interior is of great importance to understand the mechanisms of regional and global climatic changes, manage the environment, and organize societal and economic activities in the northwestern Asia. On a regional scale, the evolution of aridity in the Asian interior is probably linked to the uplift of the Tibetan Plateau (Raymo et al., 1988; Kutzbach et al., 1993; An et al., 2001), because a high plateau would obstruct the transport of moisture into the Asian interior (Boos and Kuang, 2010; Molnar et al., 2010). The study of aridification history in the Asian interior may, therefore, offer a method to constrain the growth process of the Tibetan Plateau.

On the global scale, aridity in this region may be linked to the development of glaciers in the northern hemisphere, because a colder polar region will make the westerly jet expand toward south (Broecker, 1994; Ding et al., 2005). Thus drying of inland Asia may also bear a clue to the evolution of the northern hemisphere ice volume.

The Tarim Basin is an extremely arid region in northwest China. Cenozoic ostracoda research suggested a cold and dry climate with associated humid climate in the northern part of the Tarim Basin in the late Miocene, but enhanced aridity during the Pliocene led to the shrinkage of lakes and absence of ostracoda (Sun et al., 1999). Cenozoic pollen sequences also show that the entire Tarim Basin rapidly became more arid in the Pliocene (Wang and Yan, 1987; Yan and Mu, 1990). The onset of the eolian dune and multiple climatic proxies of the late Cenozoic deposits in Sanju in the southern Tarim Basin revealed abrupt paleoenvironment changes after 5.3 Ma, with increased aridity and episodic accumulation of eolian dust in the windward slope of the Kunlun Mountains (Sun and Liu, 2006; Sun et al., 2008). In situ stratigraphic evidence for the onset of eolian dune building in the central Taklimakan Desert, shows that the eolian sand dunes developed as early as 7 Ma, suggesting a significant desertification in this area since the late Miocene (Sun et al., 2009). These recent progresses have notably increased our understanding of the aridity and desert development in the Tarim Basin since the late Miocene.

In order to further refine the aridification history of the Tarim Basin and explore its dynamic linkage with Tibetan Plateau uplift and global climate changes, a drilling project was conducted in Lop Nor in the eastern Tarim Basin in 2004. We obtained a continuous long (1050.60 m, drilling cores Ls2) core mainly consisting of fluviolacustrine sediments with associated eolian deposits. In this study, we report the high-resolution magnetostratigraphy and magnetic susceptibility of this core, aiming to provide new evidence for the late Cenozoic development of aridification in the Tarim Basin.

2. General setting and sampling

The terrestrial Tarim Basin in the northwestern China developed as a complex basin in the Cenozoic in association with mountain building in the northern Tibetan Plateau to the south and the Tian Shan to the north (Wang and Yan, 1987; Hao et al., 2002; Bosboom et al., 2011) (Fig. 1). The basin is surrounded by high mountain
ranges: the Tian Shan to the north, the Pamir to the west, and the Kunlun Mountains and Altyk Tagh to the south. The elevations of these mountains exceed 4000 m, whereas the elevation of the basin ranges from 800 to 1300 m. Glacier-fed streams transport clastic sediment from these high mountains to the basin, and wind-sorting of the glaciofluvial sediment leads to the redistribution of gravel along the edge of the basin where sand is confined to the Taklimakan Desert in the center of the basin. The Taklimakan Desert covers an area of 337,000 km², 85% of which is covered by active sand dunes. The mean annual precipitation and evaporation are <50 mm and about 3000 mm respectively, which makes it the driest desert in the world (Zhu et al., 1980).

The regional climate in the Tarim Basin is mainly controlled by westerly winds, although it is also influenced by the Asian monsoon, which reaches today its western limit on the Tian Shan (Watts, 1969). High-resolution palaeoenvironmental changes in this region since the late Pleistocene have been retrieved from a few lakes including the Lakes Chaiwopu, Manas and Balikun along the northern slope of the Tian Shan (Wang and Jiao, 1989; Han and Yuan, 1990; Rhodes et al., 1996), and Lake Lop Nor in the eastern Tarim Basin (Luo et al., 2009). However, high-resolution data constraining the longer-term climatic change in western region remain rare.

The core Ls2 was drilled near Luobuzhuang in Lake Taitema, a secondary lake of Lop Nor in the eastern Tarim Basin, in the summer and fall of 2004. The directional sensor was used to confirm the direction of the well every 100 m drilling depth and at the bottom. The core is basically vertical, with apical angles within 1–2° from the surface to the bottom. Strata bedding is basically perpendicular to the central axis of the core when the cores were cut. The depth of Ls2 is 1050.60 m, and its core recovery is 96%. According to mineral composition, grain-size, bedding, color, and structure feature, the core of Ls2 can be divided into four lithologic units from the bottom to the top. (I) 1050.60–800.68 m: grey and bluish grey argillaceous limestone intercalated with brown and bluish-grey clay. The argillaceous limestone is laminated, and the clay is layered strata. Bedding is horizontal or wavy. (II) 800.68–710.50 m: grey, bluish-grey and brown clayey silt intercalated with brown and grey clay. Again bedding is horizontal or wavy, with layered or massive structure. (III) 710.50–34.76 m: grey, bluish-grey and brown-red clayey silt intercalated with brownish red clay, and bluish grey and grey silt. Bedding is parallel, with layered or massive structure. (IV) 34.76–0 m: grey massive sand in the upper part of the drill core.

After the core was split and cleaned, U-channel samples (U-shaped, 2 cm × 2 cm square cross-section, 1.5 m in length, within non-magnetic plastic tubing, and with an arrow showing an “up” direction) were taken from one half of the split core for continuous long-core magnetic measurements. In total, 706 U-channel cores (1050 m) were taken from the Ls2 core. In addition, 236 discrete samples (2 cm × 2 cm × 2 cm) at 50–80 cm interval were taken from the top of the Ls2 core (0.13–124 m and 158.76–175.22 m).

3. Methods

Remanence was measured using a 2G cryogenic superconducting magnetometer (model 755R) housed in the magnetic shielded space (~150 nT) at the Institute of Earth Environment, Chinese Academy of Sciences. All the 706 U-channel cores were subjected to stepwise alternating field (AF) demagnetization at field up to 80 mT with 5 or 10 mT increments (measuring space is 5 cm). All the 236 discrete samples were subjected to stepwise thermal demagnetization using a TD-48 thermal demagnetizer. They were stepwise heated to 680 °C with temperature increments of 10–50 °C. The principal components direction was computed using a “least-squares fitting” technique (Kirschvink, 1980).

Low-field magnetic susceptibility (χ, calculated on a mass-specific basis) of the powder samples at 20 cm interval was measured with Bartington MS2 meter at a frequency of 470 Hz. Temperature-dependent susceptibility (χ–T) curves of representative samples were measured in an argon atmosphere with AGICO MF1-KA model kappa bridge with CS-3/L system. Temperature-dependent magnetization (J–T) curves were measured in air using the NMB-89 Magnetic Balance in the Paleo and Rock magnetism Laboratory at Center for Advanced Marine Core Research in Japan. The inducing field was 0.5 T and the heating rate was 8 °C/min. The samples were heated up to 700 °C.

4. Results and discussion

4.1. Magnetostratigraphy

Most samples yielded a stable characteristic remnant magnetization (ChRM) component after stepwise AF demagnetization up to 80 mT or thermal demagnetization up to 580 °C, which indicates that magnetite is the dominant carrier of ChRM (Fig. 2). However, some samples had to be heated to 680 °C in order to determine a stable ChRM component. This suggests the presence of hematite. More than three successive points in the orthogonal diagrams were used to determine the direction of ChRM during the establishment of polarity sequence. Finally, the isolated ChRM vector directions with
maximum angular deviations (MAD) smaller than 10° and inclination ranging from 15° to 75° were considered as acceptable data. 16,685 AF demagnetization results were used to calculate ChRM. Among the 236 discrete samples, 194 (82%) samples gave reliable ChRM directions. The AF and thermal demagnetizations resulted in an overall consistent change in the ChRM vector directions (Fig. 2), except for some short intervals, possibly because the 80-mT AF demagnetization is inadequate to isolate a stable ChRM that is carried by hematite. The detailed geomagnetic polarity variations of the Ls2 core samples were established based mainly on the AF demagnetization isolated ChRM vector directions, but with the consideration of thermal demagnetization data. Results suggest that the Ls2 core recorded 14 normal (N1–N14) and 13 reversal (R1–R13) polarity zones (Fig. 3).

Since surface water existed in the Lake Taitema until the early 1980s (Fan, 1987; Zhong et al., 2005), it was likely that the succession studied here terminated in late Quaternary. As shown in Fig. 3, in the top of the Ls2 core, local magnetozone N1 (at depth 0–36.7 m) should thus correlate to the chron C1n of CK95 GPTS (Cande and Kent, 1995), consistent with the other researches concerning the age model in the Lake Taitema (Fan, 1987; Zhong et al., 2005). In the upper part of the Ls2 core (at depth 36.70–224.60), the distinctive reversal polarity with three normal polarities (two are short and one is long) (R1–R4) appear to correlate to chron C2r.2r to C1r.1r (Matuyama chron) of CK95 GPTS. The distinctive magnetozones (N7–N5) between 224.60 and 439.45 m in the Ls2 core are characterized by three normal and two reversals that correlate with the Gauss chron.

The boundary of B/M, M/G and G/G are in 36.70 m, 224.60 m and 439.45 m, respectively. The Ls2 core is estimated to cover an age from 7.1 Ma to the present. The boundaries of the Miocene-Pliocene (5.332 Ma) and the Pliocene-Pleistocene (2.588 Ma) (Ogg, 2009) are located at 746.7 and 225.6 m depths of the Ls2 core, respectively.

The diagram of stratigraphic depth versus magnetostratigraphic age of Ls2 shows an abrupt decrease in the rate of sedimentation accumulation at 1.77 Ma (Fig. 4). This abrupt change is associated with a decrease in the accommodation space for sediment accumulation, which was possibly triggered by tectonic movement. A similar decrease in sedimentation rate was observed in Atushi on the western margin of the Tarim Basin (Heermance et al., 2007). And the deformation and same dips of the Pliocene and late Miocene sediment in the Maza Tagh (Sun et al., 2009) suggest that the tectonic deformation has occurred in the Tarim basin after Pliocene. There are two more modest increases in accumulation rate from 120.16 m/myr (7.07–6.567 Ma).
to 207.61 m/m.y. (6.567–5.894 Ma) and from 177.36 m/m.y. (5.894–3.58 Ma) to 271.76 m/m.y. (3.58–3.04 Ma). These two changes in sedimentation rate correlate approximately with two main pulses of sedimentation in the Tarim Basin at 7–6 Ma and 3.6–3 Ma estimated from an entirely independent method of balanced section reconstruction (Métivier and Gaudemer, 1997) and from different areas in the Tarim Basin (Zheng et al., 2000; Sun and Liu, 2006; Heermance et al., 2007). They also appear to be typical of sedimentation rate in foreland-basin environments (100–400 m/m.y.) (Burbank et al., 1992; Harrison et al., 1993).

4.2. Magnetic mineralogy

χ–T curves are highly sensitive to mineralogical changes during thermal treatment, but such changes can provide important information about magnetic mineral composition (Dunlop and Özdemir, 1997; Deng et al., 2004; Ao et al., 2009). Fig. 5 illustrates the χ–T curves of the representative samples from drilling core Ls2 in the eastern Tarim Basin. Among these selected samples, two (h–i) are selected from unit I, two (f–g) from Unit II, three (c–e) from unit III, and two (a–b) from unit IV. All the χ–T heating curves show a significant increase above ~400 °C,
a peak at ~510 °C, and a major decrease in magnetic susceptibility after ca. 580 °C. Although the increased magnetic susceptibility is remarkable, the magnetic susceptibility is high before 400 °C than that after 580 °C. This suggests the existence of magnetite in the sediments, other evidences also support this result, as discussed later on. Most samples have an increased $\chi$ cooling after heating to 700 °C, and the noteworthy peak at about 510 °C not seen in the cooling curves. This is interpreted to result possibly from neo-formation of magnetite via annealing of iron-containing paramagnetic minerals (e.g., chlorite) (Deng et al., 2005; Ao et al., 2009).

$J$–$T$ curves allow for determination of the Curie point of magnetic minerals and an estimation of their thermal stability (Grommé et al., 1969; Torii et al., 2001), which is a useful tool for identification of remanence carriers. In order to further examine the magnetic minerals present in the deposits of Ls2 core, 9 samples were chosen for measurements of J–T curves (Fig. 6). J–T curves of samples in unit IV (a–b) show a clear Curie temperature at ca. 580 °C, consistent with the dominance of magnetite in unit IV. The J–T curves of samples in unit III (c–e) show Curie temperatures of 580 °C and 650–670 °C, which indicates that both hematite and magnetite are important in these sediments. The J–T curves of samples in unit II (f–g) and I (h–i) are all characterized by Curie temperature of 110 °C and 580 °C, which suggests that magnetite and goethite are the main magnetic mineral in those two parts of Ls2.

Magnetic susceptibility is a measure of the concentration of ferromagnetic minerals in the sample, which is closely related to the sediment provenance and/or climate changes (An et al., 1991, 2001; Maher and Thompson, 1991; Dearing et al., 1998; Guerrero et al., 2000; Ao et al., 2010). Variations of the magnetic susceptibility of Ls2 core show consistent changes with the lithology (Fig. 7). In the lowest part (unit I, 7.1 Ma to 5.6 Ma, at depth 800.68–1050.60 m), the magnetic susceptibility shows the largest amplitude variation (from 2 to 58×10$^{-8}$ m$^3$ kg$^{-1}$). In the second part (unit II, 5.6 Ma to 5.1 Ma, at depth of 710.50–800.68 m), magnetic susceptibility gradually changes, but with two peaks. In the third part (unit III, from 5.1 Ma to 0.64 Ma, at depth 30.02–710.50 m), the magnetic susceptibility is relatively stable and generally lower than 20×10$^{-8}$ m$^3$ kg$^{-1}$. In the fourth part (unit IV from 0.64 Ma–present, at depth 0–30.02 m), magnetic susceptibility is similar to that in unit I.

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 microbial populations on the one hand and the sedimentation rate on the other (Karlin and Levi, 1983; Snowball, 1993). However, the organic carbon content is quite low (<0.8%) in the sedimentary sequence.
drilled from Lop Nor (unpublished data) and the sedimentation rate is very high (ca. 120–208 m/Ma). So, perhaps the magnetic minerals in the sediments from Lop Nor were not significantly affected by post-depositional dissolution.

One possible cause of changes in the concentration of magnetic mineral in the Ls2 drilling core may be the variations in abundance of magnetic minerals in the catchments, which is directly related to the intensity of pedogenesis, with strong pedogenesis creating increasing pedogenic ferrimagnetic minerals, and weak pedogenesis bringing in decreasing pedogenic ferrimagnetic minerals (Ao, 2010). There are two critical stages in the process of secondary ferrimagnetic mineral formation that involve the supply of Fe and secondly transformation from ferrihydrite–maghemite–hematite (Barrón and Torrent, 2002) and ferrihydrite–magnetite (Dearing et al., 1996). The role of climate largely through rainfall would argue strongly for the ferrihydrite–magnetite transformation to be dominant in temperate regions (Blundell et al., 2009). Torrent et al. (2006) characterized samples from soil and paleosols in world region from Russian steepe to Chinese Loess Plateau to Argentinian Pampa loess. They suggested that transformation from maghemite to hematite is fast than from ferrihydrite to maghemite in areas with lower rainfall and longer dry seasons. So, maghemite is little or no occurrence in this type region. The result of rock magnetism research in the Taklimakan desert in the Tarim basin showing that magnetite is dominant magnetic mineral, and a minor high-coercivity mineral (possibly goethite) (Torii et al., 2001) are consistent with this inference. For the Chinese loess-paleosol sequence, strong pedogenesis has produced a large number of ferrimagnetic minerals and significantly enhanced the magnetic susceptibility in the paleosol layers (e.g. Zhou et al., 1990).

The pedogenesis in the Tarim Basin is closely related to westerly precipitation, with strong precipitation creating enhanced pedogenesis, and weak precipitation leading to weakened pedogenesis. Thus the decreased magnetic susceptibility after ca. 5.1 Ma may suggest weakened pedogenesis and increased aridity and decreased westerly precipitation in the Tarim Basin after this time. This timing is consistent with the paleoclimate inferences based on paleontology, pollen, and the onset of dunes from other late Cenozoic sections in the Tarim Basin (Wang and Yan, 1987; Yan and Mu, 1990; Sun et al., 1999; Sun and Liu, 2006; Sun et al., 2008). However, it disagrees with the onset of eolian dune formation in Maza Tagh in the central Tarim Basin (Sun et al., 2009). We ascribe this earlier onset of aridity to an erroneous correlation of the magnetostratigraphy to the geomagnetic reversal history or to spatial differences in climate change in the Tarim Basin. We tentatively offer a re-interpretation of the magnetostratigraphic results by Sun et al. (2009), which indicates an age of 6 Ma for the bottom of Maza Tagh section (Fig. 8). Although a vertebrate fossil of Olonbulakia tsaidamensis was discovered at the depth of 885 m (re-interpreted age is 5.3 Ma) in Maza Tagh section, which was suggested to have an age of 8–9 Ma (Deng, 2006). In any case, it shows only that the age of this fossil-bearing layer is younger than or equal to the fossil age. Moreover this sole fossil was possibly excavated by fluvial erosion and re-deposited in that layer at

![Fig. 6. J-T curves of representative samples from drilling core Ls2.](image-url)
5.3 Ma. Hence this new age suggests that the fossil bearing layer is younger than the assigned fossil age, the fossil may have been excavated from older sedimentary rock by a river and re-deposited in that layer after 5.3 Ma.

Tectonic movements in the provenance area (Storti and McClay, 1995) and frequent climatic changes (Molnar and England, 1990; Zhang et al., 2001) should increase the sedimentation rate. And the sedimentation rate in Maza Tagh section should have increased since the Pliocene because the provenances (Kunlun Mountains) were uplifted at about 5.3 Ma (Sun and Liu, 2006). The re-interpretation of the data shows that sedimentation rate increased in the early Pliocene; but the original interpretation disagrees with that result (Fig. 8).

Our results show that rapid aridification enhancement in the Tarim Basin began at 5.6 Ma, followed by two short intervals of less arid climate during 5.6–5.1 Ma. Aridity is the prevailing climatic feature in this region since 5.1 Ma. The increase in magnetic susceptibility in Ls2 is generally accompanied by grain size coarsening of sediments in late Quaternary. It is the result of abundant melt-water-fed rivers, as well as at other sites in the northern Tarim Basin (Sun et al., 1999) and the Qaidam Basin (Wang et al., 2007).
4.3. Possible driving mechanisms of the Asian aridity at the Miocene/Pliocene boundary

Two major factors are possibly responsible for the enhanced aridity in the Tarim Basin at the Miocene/Pliocene boundary: northern hemisphere cooling and Tibetan Plateau uplift. During the late Miocene and early Pliocene, rapidly enhanced cooling in the northern hemisphere (Zachos et al., 2001), possibly due to Mediterranean desiccation (Adams et al., 1977), would have enabled sea-ice formation and global glacial expansion with more extensive ice cover at higher latitudes (Hodell et al., 1986). This would influence the pressure systems in North Asia such as the Mongolia high and Siberia high (Porter and An, 1995). The intensification of these two systems and southward shift of the polar front would in turn result in a southward migration of westerly zone (Luo et al., 2009). The intensified westerly winds should facilitate moisture transport and bring rainfall to the southern Tibetan Plateau, but decrease the precipitation in the northern Tibetan Plateau, especially the Tarim Basin (Liu and Yin, 2001). Therefore, this process could result in enhanced aridity in the Tarim Basin. The massive floods in the southern area during that period (Mckenzie, 1999) are consistent with the migration of westerly precipitation towards the south.

In addition to the northern hemisphere cooling, the growth of the Tibetan Plateau can influence the climatic changes in Asia as well (Kutzbach et al., 1989; Ruddiman et al., 1989; Raymo and Ruddiman, 1992; An et al., 2001; Liu et al., 2003; Zhang et al., 2007). Climate-model simulations show that continued uplift and expansion of the plateau along its northern and eastern margins should modify Asian atmospheric circulation patterns, and increase the aridification in central Asia (An et al., 2001). Although the timing and height of the Tibetan Plateau uplift are still controversial, more evidence from both the interior of the Tibetan Plateau and surrounding basins indicates that large amplitude uplift occurred in late Miocene (Tapponnier et al., 2001; Molnar, 2005; Sun and Liu, 2006). The Tibetan Plateau blocks more effectively the penetration of moisture from the south Asian monsoon into Tarim Basin and thus enhances the rain shadow effect (Sun et al., 2008; Boos and Kuang, 2010; Molnar et al., 2010).

5. Conclusions

1. Magnetostratigraphic results show that the Late Cenozoic sedimentary sequence drilled from Lop Nor in the eastern Tarim Basin has recorded a magnetic polarity from C3Bn to C1n, covering the last 7.1 Ma.
2. Decreased magnetic susceptibility after ca. 5.6–5.1 Ma indicates an enhancement of aridity in the Tarim Basin since the Miocene/Pliocene boundary.
3. We attribute this aridity in the Tarim Basin to the northern hemisphere cooling and the uplift of the Tibetan Plateau.

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