1 **Paleomagnetic chronology and paleoenvironmental records from drill cores from**
2 **the Hetao Basin and their implications for the formation of the Hobq Desert and**
3 **the Yellow River**

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Abstract

Reconstructing the Cenozoic environmental history of Hetao Basin, in the northern part of the Ordos Plateau in North China, is important not only for revealing the evolution of the Yellow River, but also for understanding the formation of the Hobq Desert. Here we present the stratigraphic framework of drill core DR01 with length of 2503.18 m, and the results of magnetostratigraphic and ESR dating and multi-proxy analyses of drill core WEDP05 with length of 274.60 m, from the Hetao Basin. The magnetostratigraphic and ESR results indicate that core WEDP05 spans the last ~1.68 Ma. Stratigraphic sequence of core DR01 indicates that the Hetao area was uplifted and eroded during the early Cenozoic, before subsiding to form a sedimentary basin. Subsequently, the basin was a fluvio-lacustrine environment during the Pliocene and then experienced alternating desert and fluvio-lacustrine conditions during the Quaternary. Sedimentary facies and multi environmental-proxy analyses of core WEDP05 indicate that the basin was occupied by a fluvio-lacustrine system during the following intervals: ~1.47 - ~1.30 Ma, ~1.17 - ~1.07 Ma, ~0.68 - ~0.60 Ma and from ~0.47 Ma to the last interglacial; and that a desert environment developed during the lake regression phases of ~1.30 - ~1.17 Ma, ~1.07 - ~0.68 Ma and ~0.60 - ~0.47 Ma. The presence of aeolian sand at the base of core WEDP05 suggests that the origin of the Hobq Desert can be traced back to the early Pleistocene, and resulted from the erosion and transportation of exposed fluvio-lacustrine sediments by near-surface winds associated with the Asian winter monsoon. A large river channel in the Hetao Basin may have existed as early as the Pliocene, which was

35 occupied by the Yellow River when its upper reaches formed by at least ~1.6 Ma.

36 Subsequently, at least since ~1.2 Ma, the Yellow River formed its drainage system

37 around the Hetao Basin and controlled the paleoenvironment evolution of the basin.

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39 **Key words:** the Hetao Basin; paleoenvironment; the Hobq (Kubuqi) Desert; the

40 Yellow River

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1. Introduction

The Hetao Basin is located in the upper reaches of the Yellow River and in the transitional zone between the monsoonal-influenced Loess Plateau to the south and east; and arid areas, including gobi and sand deserts, to the north and west (Fig. 1). The stratigraphy and chronology of the sedimentary deposits of the basin are crucial not only for revealing the evolution history of the Yellow River, but also for understanding the formation of the Hobq Desert. The formation and evolution of the deserts in northwest China are topical in paleoenvironmental research because they are closely associated with the processes of aridification of the Asian interior, the uplift of the Tibetan Plateau and global cooling during the Cenozoic (e.g. Sun and Liu, 2006; Sun et al., 2011a; Yang et al., 2012; Z. Li et al., 2014; Wang et al., 2015; Zheng et al., 2015). Evidence from the marginal and central Tarim Basin has suggested that the age of origin of the Taklimakan Desert was ~3.4 Ma (Sun et al., 2011a), ~5.3 or ~5.1 Ma (Zheng et al., 2000; Zheng et al., 2003; Sun and Liu, 2006; Chang et al., 2012), ~7 Ma (Sun et al., 2009), or even between ~26.7 Ma and 22.6 Ma (Zheng et al., 2015). Recent chronological studies of deserts proximal to the Loess Plateau indicate that the Badain Jaran Desert formed at ~1.1 Ma (Wang et al., 2015) and the Tengger Desert at ~0.9 Ma (Z. Li et al., 2014). Among these deserts, the proximal deserts of the Loess Plateau are the more important, because 1) they are located upwind of the Loess Plateau, beneath the pathway of the Asian winter monsoon, which is the major dust transport agent (e.g. Liu, 1985; An et al., 1991); and 2) studies of isotope geochemistry suggested the proximal deserts are an important source of the aeolian

materials supplied to the downwind areas in eastern Asia (e.g. Chen et al., 2007; G. Li et al., 2007; Rao et al., 2008).

The Hobq Desert is located in the southern part of the Hetao Basin, to the north of the area enclosed by the great northern bend of the Yellow River (Fig. 1B). It is one of the main proximal deserts of the Loess Plateau (Sun, 2002; Rao et al., 2008). Its formation was controlled by the evolution of the Yellow River, and is also closely related to the origin of the Loess Plateau. Previous research on the evolution of the Hobq Desert mainly focused on the age of the modern desert landscape (Yan, 2004; J. Sun et al., 2006; S. Li et al., 2007; Fan et al., 2013), and several studies have proposed that human activity over the past several thousand years played an important role in the desertification of the area (Wang, 1991; Yan, 2004). The chronology of several excavated profiles in the desert, with a depth of only several meters, revealed that the active sand dunes in the Hobq Desert had formation ages ranging from the late Pleistocene to the Holocene (J. Sun et al., 2006; S. Li et al., 2007; Fan et al., 2013). These findings have been interpreted as indicating that the dunes were formed as a result of both climatic change and human activity (J. Sun et al., 2006; Fan et al., 2013). A recent study of the dunes directly overlaying bedrock suggested a Holocene age for the formation of the sand sea in the Hobq Desert, which was interpreted as resulting from sediment supply from both the Yellow River and as a result of human activity (Yang et al., 2016). Despite these findings, the longer-term history of the Hobq Desert remains unclear because complete and representative sedimentary outcrops are sparse in the central part of the desert due to coverage by sand dunes.

As a formation model for the Hobq Desert, several authors have proposed that the aeolian sand of the desert was mainly transported by the Asian winter monsoon from the adjacent banks of the Yellow River (Liu et al., 2005; Rao et al., 2008; Rao et al., 2011; Nie et al., 2015; Yang et al., 2016). The evolution of the Yellow River not only supplied sand material to the Hobq Desert downwind, but it also had a significant impact on the sedimentary environment of the Hobq Desert due to intermittent flooding into the adjoining desert area, resulting in the formation of a lake environment. Thus, the evolution of the Yellow River in the Hetao Basin had a significant impact on the paleoenvironmental evolution of the Hobq Desert, which in turn is closely related to the paleoenvironment of the Hetao Basin.

Investigations of lacustrine strata in outcrops and drill cores from sites surrounding the Hetao Basin have revealed that a lake covered the basin in the late Pleistocene (Jia et al., 2001; Ma et al., 2004; Li et al., 2005; Deng et al., 2007; Chen et al., 2008; Jiang et al., 2013; Li et al., 2015; Jia et al., 2016). In addition, regional geological investigations and many oil exploration drilling records have revealed that lake deposits were widely developed in the Hetao Basin during the Quaternary (Zhang et al., 1983; China Earthquake Administration (CEA), 1988). These results suggest that this lake environment may have existed intermittently since the early Pleistocene. Unfortunately, the poor age constraints in this work prevent us from better understanding the paleoenvironmental evolution of the area. With regard to the evolution of the Yellow River, geomorphological work suggested that its upper reaches were formed in the early Pleistocene (Li, 1991; Li et al., 1996, 1997;

Craddock et al., 2010), as a result of tectonic (Li, 1991; Li et al., 1996, 1997) or climatic factors (Craddock et al., 2010). In the case of the middle reaches of the Yellow River, studies of river terraces and geological evidence suggested that they were formed in the late Miocene-early Pliocene (Li and Yang, 1991; Lin et al., 2001) or late Pliocene (Nie et al., 2015); whereas others have argued that they were initially formed in the early Pleistocene (Cheng et al., 2002; Pan et al., 2005; Hu et al., 2012; Pan et al., 2012; Hu et al., 2016) or late Pleistocene (Wang et al., 2002; Jiang et al., 2007; Prins et al., 2009; Zhang et al., 2009; Guo et al., 2012; Liang et al., 2015). With regard to their origin, several studies suggested that they are a result of faulting and rifting around the Ordos Block (e.g., Lin et al., 2001); whereas others have advocated that they were formed by the interconnection of isolated reaches or paleolakes in basins along the course of the Yellow River (Li and Yang, 1991; Jiang et al., 2007; Qiu et al., 2008; Zhang et al., 2009; Pan et al., 2012; Liang et al., 2015). From the foregoing, it is clear that study of a well-dated stratigraphic sequence within the Hetao Basin near the Hobq Desert is needed in order to better understand both the paleoenvironmental history of the Hetao region, and the formation processes of the Yellow River.

In the present paper, the stratigraphic sequence of core DR01 from the Hetao Basin, spanning the entire Cenozoic, is described; and the results of magnetostratigraphic and ESR dating and multi-proxy analyses of core WEDP05, from the centre of the basin and spanning the Pleistocene, are presented. The results are used to discuss the paleoenvironmental history of the Hetao Basin, the age of

formation of the Hobq Desert, and the evolution of the drainage of the Yellow River within and around the Hetao Basin.

2. Geological and tectonic background

The Hetao Basin, the largest fault-subsiding basin on the periphery of the Ordos Block (Fig.1B), is located in the central Inner Mongolia Autonomous Region, China, and exhibits an east-west zonal distribution pattern with an average altitude of 900-1200 m. It is bordered by the Ordos Plateau to the south, the Yinshan Mountains to the north, and the Langshan Mountains to the northwest (Fig.1C). Due to the tectonic uplift of the Yinshan Mountains and the Ordos Plateau, and associated erosional process from the late Cretaceous to early Cenozoic, the Hetao area subsided and formed a sedimentary basin during the middle Cenozoic (CEA, 1988). As a result of regional tectonic activity, a series of faults developed in the border areas between the Hetao Basin, the Yinshan Mountains and the Ordos Plateau (Fig.1C). Normal faults are distributed mainly along the southern piedmont of the Yinshan Mountains, and buried faults mainly along the southern margin of the basin, and within the basin itself (CEA, 1988). The faults divide the basin into four depressions (the Jilantai, Linhe, Baiyanhua and Hohhot Depressions) and three uplifted areas (the Dengkou, Xishanzui and Baotou Uplifts) (Fig.1C).

Fig. 1

Geological survey has revealed that the basement of the basin is Precambrian granitic gneiss, which is unconformably overlain by the Cretaceous and Cenozoic sedimentary sequence (Zhang et al., 1983; CEA, 1988; Cai et al., 1988; Fu et al.,

1994). The Cretaceous strata are mainly composed of lacustrine reddish-brown, purplish red and grey clay and silty clay in the lower part; and grey black clay, silty clay interbedded with brown clay and silty clay in the upper part (Cai et al., 1988). The Cenozoic strata within the Hetao Basin are well-developed with a thickness of over 5000 m (CEA, 1988). The Oligocene strata consist of fluvial-lacustrine brown, green-greyish clay and silty clay interbedded with glutenite, which unconformably overlies the Cretaceous strata, and with a thickness of more than 1000 m (Zhang et al., 1983; CEA, 1988; Fu et al., 1994). The Miocene strata consist of brown clay, silty clay interbedded with grey silty clay and sand, with a thickness of more than 1200 m (CEA, 1988). The Pliocene strata are characterized by fluvial-lacustrine brown, yellow-greyish clay and silty clay interbedded with yellow-greyish sand, with a thickness of more than 900 m (CEA, 1988). The Quaternary sediments are characterized by thick grey and green-greyish mudstone and yellow sandstone formed in fluvio-lacustrine environments, with thicknesses ranging from 0-2500 m (CEA, 1988). For example, in the Linhe Depression (Fig. 1C), the depocenter of the basin during the Quaternary, the Quaternary strata are more than 2400 m thick; while in the Baotou Uplift (Fig. 1C), the thickness is less than 200 m and the basement gneiss can be observed directly in some outcrops (CEA, 1988).

Meteorologically, the Hetao Basin is located in the transition zone between arid and semi-arid regions, on the northern periphery of the area of influence of the East Asian summer monsoon (Fig. 1A). The annual average temperature is about 4-8 °C and the mean annual precipitation about 275 mm. About 60-70% of the dust transport

takes place in the spring and winter seasons (Wang et al., 2003). As a consequence of the dry and windy environment, together with available clastic material originating from the Yellow River, the Hobq Desert and the Ulan Buh Desert formed in the west and southwest parts of the basin, respectively (Fig.1C).

The Hobq Desert is located on the transitional slope between the northern Ordos Plateau and the Hetao Basin (Fig.1C). Active sand dunes comprise more than 80% of the total (Zhu et al., 1980; Rao et al., 2011; Yang et al., 2012) and are distributed in the west and north parts of the desert, while semi-fixed and fixed sand dunes are mainly in the south and east parts. The mean annual temperature is 5 °C to 7 °C, the mean annual precipitation ranges from 240 mm in the northwest to 360 mm in the southeast and the mean annual evaporation is 2000-2700 mm (Rao et al., 2011; Liu et al., 2016; Yang et al., 2016).

3. Methods

3.1. Drilling and sampling

From July to December, 2009, a geothermal drill core (DR01) (40°40'27.09"N, 108°18'29.13"E, Fig. 1C) was obtained by the Inner Mongolia Second Hydrogeology Engineering Geological Prospecting Institute. The core reached a depth of 2503.18 m and penetrated the entire sequence of Cenozoic sediments, reaching the underlying Cretaceous strata. The stratigraphy of the Hetao Basin since the Cretaceous has been studied intensively and the detailed petrographic descriptions have been presented (Zhang et al., 1983; CEA, 1988; Cai, 1990; Fu et al., 1994). Based on the regional stratigraphical correlations, a rough chronological framework of drill core DR01 can

be established and was presented in Xu et al. (2012).

From March to May, 2013, drilling was conducted in the centre of the basin by the “West Environment Drilling Project, China (WEDP)”, which focused mainly on the evolution of the deserts in northwestern China, using an XY-4 drilling rig. The wall of the hole was consolidated with casing pipes from 0 to 50 m and with a mud protection from 50 m to the bottom of the core. Core WEDP05 (40°45'9.04"N, 108°23'31.78"E) is located at the junction zone of the Yellow River and the Hobq Desert (Fig. 1C). The length of the core is 274.6 m with a mean core recovery rate of 72%. The recovered interval consisted mainly of finer-grained sediments such as clay, silt clay, and fine sand, while coarse sand was incompletely recovered.

The uncompacted parts of core WEDP05 were cut into 2-cm-thick slices in the field. The compacted core sections were transported to the laboratory where they were first cut into 2-cm-thick slices perpendicular to the core axis. Subsequently, samples of 2 cm³ volume were cut for paleomagnetic measurements and the remaining material was retained in the form of powder samples for other analyses. In addition, eight samples for electron spin resonance (ESR) dating were collected from the central portion of the sediment cores immediately after they were obtained. The samples were collected using steel tubes inside an opaque shield, which were then covered and sealed in black plastic bags to avoid light exposure and changes in water content during transportation.

3.2. Chronology of core WEDP05

497 paleomagnetic samples were obtained from core WEDP05 at an interval of

20-50 cm for demagnetization using a MMTD80 thermal demagnetizer. The remanent magnetization was measured using a 2G Enterprise 760 Cryogenic Magnetometer. Both the demagnetizer and the magnetometer are installed in a shielded room with a residual field of less than 200 nT. The measurements were made in the Key Laboratory of Western China's Environmental Systems (Ministry of Education), Lanzhou University. The samples were demagnetized at steps of 40°C from room temperature to 600 °C or 680°C and the results evaluated using orthogonal projection plots. Characteristic Remanent Magnetization (ChRM) directions were calculated using principal component analysis (PCA) (Kirschvink, 1980). The paleomagnetic data were analyzed using PMGSC (V4.2) developed by R. Enkin, the least-squares linear fits were anchored to the origin on the diagrams.

For the ESR dating, the eight samples from core WEDP05 were dried at low temperature, and the mass of sample was measured before and after drying to determine the water content. The dried samples were disaggregated and sieved to isolate the 125-200 mm fraction and were then treated with 30% H₂O₂ to remove organic matter and then with 40% HCl for 24 h to remove carbonates. The samples were then washed with distilled water and etched with 40% HF for 40 min to remove feldspar grains and the alpha irradiated edge of the grains. After rinsing to neutral with distilled water, magnetic minerals were removed by magnetic separation. After rinsing with distilled water and drying at low temperature, the samples were examined under a microscope to check that pure quartz grains had been obtained. The extracted quartz grains were then split into 5 aliquots, each of 120 mg. The aliquots were exposed to a

^{60}Co irradiation field to receive various additional gamma doses, which were monitored by the standard alanine dosimeter with an error of 3.2%. The dose rate was calculated from the concentrations of uranium, thorium, and potassium of each sample (Adamiec and Aitken, 1998), the water content was taken into account in order to correct the dose rate. The ESR spectra of the quartz E' centre was analyzed using an ER-200D-SRC electronic spin resonance spectrometer with microwave frequency 9.7652 GHz, microwave power 0.2 mW, modulation amplitude 0.25 G, modulation frequency 100 GHz and time constant of 50 ms. The pretreatment and measurement of the samples were performed in the Applied Nuclear Technology Institute of Chengdu University of Technology.

3.3. Measurement of environmental proxies

938 powder samples were collected from core WEDP05 at a 20-cm interval for environmental proxy index measurements. For measurements of sediment color, the samples were air-dried and then disaggregated and passed through a 40 μm mesh sieve. The color values of L^* , a^* and b^* were measured using a Konica-Minolta CM700d color meter. In this color space, L^* indicates lightness and ranges from 0 to 100; a^* is green (negative) to red (positive); and b^* is blue (negative) to yellow (positive). For grain-size measurements, the samples were pretreated using 10% H_2O_2 to remove organic matter and 10% HCl to remove carbonates. After standing for 24 h, the residues were dispersed with 10% $\text{Na}(\text{PO}_3)_6$ and ultrasonicated for 10 min. The grain-size measurements were made used a Malvern Mastersizer 2000 laser grain-size analyzer with a measurement range of 0.02-2000 μm and a 0.1 Φ resolution. All of

above measurements were made in the Key Laboratory of Western China's Environment Systems (Ministry of Education), Lanzhou University.

Additionally, magnetic hysteresis loops and thermomagnetic curves of 20 representative samples from core WEDP05 were measured in the State Key Laboratory of Estuarine and Coastal Research, East China Normal University, using a variable field translation balance in order to investigate the rock magnetic properties of the sediments. 15 representative samples were also selected for scanning electron microscope (SEM) analysis. After removal of carbonate and organic matter using HCl and H₂O₂, the dispersed grains were washed with distilled water and then dried at room temperature. The grain morphology was observed on a Hitachi S4800 SEM in the Key Laboratory for Magnetism and Magnetic Materials of the Ministry of Education, Lanzhou University.

3.4. Sedimentary facies analysis

Visual observation, grain-size analysis and grain morphology analysis (SEM) were used to determine the sedimentary facies. For some deposits, the sedimentary facies can be recognized by textural observations. For example, clay with horizontal bedding and laminations is generally classified as a lacustrine deposit; and greenish sediments are generally considered to have been transformed by hydrochemical processes. However, when the sedimentary characteristics are unclear, grain morphology observation and grain-size component analysis can be used to identify the sedimentary facies.

The method of grain-size component analysis has been widely applied to

sedimentary facies determination and the principle and procedures of this approach have been described in detail (Sun et al., 2002, 2004, 2008, 2011a, 2011b, 2011c; Sun, 2004; Z. Li et al., 2014; Zhang et al., 2014; Wang et al., 2015; Wang et al., 2016). Briefly, three steps are involved, as follows.

Component identification. On grain-size frequency distribution curves, almost all of samples of the strata are poly-modal sediments. Each grain-size component of the sample can be mathematically partitioned based on the non-smoothed knee point on the frequency distribution curve of the grain-size data.

Function fitting and component partitioning. The appropriate mathematical functions were selected to fit the measured grain-size data and to calculate the parameters, namely, modal size and proportion, on the basis of goodness-of-fit criteria (Sun et al., 2002; Wang et al., 2016). In our analysis, the Normal and Weibull were the two function types used because in previous experiments they have been demonstrated to yield the best fitting results for most sediments (Sun et al., 2002).

Analysis of component origin. Based on the lithological characteristics and the fitting results, the origin of each sedimentary component of the typical sediment types in core WEDP05 was first recognized. Since the size and proportion of a sedimentary component are determined by, and thus reflect, the attributes of the transport processes, components with definite origins have specific ranges of size and proportion. Therefore, based on statistics of the frequencies of modal sizes and proportions of the components in all of the samples from the core, and on comparison with the size and proportion distributions of defined components in typical sediments,

the origin of each component can be identified. Three types of statistical plots, including the frequency of component size, the frequency distribution of component proportion, and a size-proportion plot, are used for component origin analysis.

4. Results

4.1. Stratigraphy

4.1.1. Drill core DR01

The sedimentary sequence of drill core DR01, from bottom to top, consists of lower Cretaceous, late Neogene and Quaternary strata (Fig. 2). The ages were determined by regional stratigraphical correlations (Xu et al., 2012).

The lower Cretaceous strata (2503.18-2125.45 m) are composed of reddish-brown clay interbedded with grey clay in the lower part; and reddish-brown clay and silty clay interbedded with yellowish sand and violet clay layers in the upper part.

The late Neogene strata (2125.45-1834.55 m) consist of light-brown clay, silty clay and pink, yellowish sand interbedded with occasional layers of violet clay and silty clay.

The Quaternary strata (1834.55-0 m) consist of well-sorted and rounded yellowish sand interbedded with occasional reddish silty clay and reddish, pink sand layers from 1834.55-1241.25 m; well-sorted yellowish sand with rounded grains interbedded with reddish, green-greyish clay, and silty clay from 1241.25 -180 m; and greyish clay and yellowish and greyish sand in the top 180 m.

4.1.2. Drill core WEDP05

The bedding of the strata in core WEDP05 is basically perpendicular to the central axis of the core. The sedimentary sequence can be divided into six lithological units (Units 1 to 6) from top to bottom, according to variations in grain-size, bedding, color and structure (Fig. 2).

Unit 1 (0-11.26 m): greyish coarse sand interbedded with yellowish coarse sand.

Unit 2 (11.26-51.92 m): green-greyish clay, grey clay, silty clay and fine sand.

The green-greyish clay (11.26-16.78 m) and grey clay layers (24.00-47.66 m) have conspicuous horizontal bedding (Fig. 2C).

Unit 3 (51.92-87.51 m): green-greyish silty clay with conspicuous horizontal bedding (from 53.20-59.89 m, and from 75.65-87.51 m) interbedded with yellowish silty clay layers (59.89-75.65 m).

Unit 4 (87.51-141.63 m): well-sorted, greyish and yellowish fine sand with rounded grains, with reddish and green-greyish clay layers. The green-greyish clay in the middle part of the unit (119.69-123.75 m) has horizontal bedding.

Unit 5 (141.63-183.35 m): reddish clay and silty clay with a layer of well-sorted yellowish fine sand (154.10-160.39 m).

Unit 6 (183.35-274.60 m): well-sorted yellowish fine sand interbedded with three reddish silty clay layers (206.82-217.25 m, 221.07-235.49 m and 268.14-274.60 m) and a grey fine sand layer (255.26-262.80 m).

Fig. 2

4.2. ESR dating

The ESR dates have potentially large errors and therefore they were only used to

check magnetostratigraphic chronology. The results of ESR dating of eight samples from core WEDP05 are listed in Table 1. The ESR ages of the eight samples decrease gradually upwards, ranging from ~1250 ka at 240 m to ~170 ka at 75 m, and indicate that the sediments at the top of the core are younger than 0.78 Ma, and the sediments at the bottom of the core are older than 0.78 Ma.

Table 1

4.3. Rock magnetic and paleomagnetic results

Magnetic hysteresis loops and thermomagnetic curves of representative samples from core WEDP05 (Fig. 3) were used to characterize the magnetic mineral assemblages of the sediments. For some green-greyish and grey clay and silty clay samples, their hysteresis loops resemble those of typical pseudo-single domain magnetite and close below ~400 mT (Fig. 3A). Corresponding thermomagnetic curves reveal two distinct peaks, one at ~300 °C and another at ~500°C, beyond which the magnetization decreases sharply to approach the Curie temperature of magnetite (580°C) (Fig. 3B). Likewise, the cooling curve increases below 580 °C, and is higher than that of the heating curve (Fig. 3B), which indicates the neoformation of magnetite during the thermal treatment. The shape of the thermomagnetic curves described above is indicative of greigite (e.g. Roberts, 1995; Roberts, et al., 2011), suggesting that greigite is the dominant magnetic carrier in these green-greyish and grey clay and silty clay samples. For the reddish clay and silty clay samples, the hysteresis loops have a wasp-waisted shape and are not saturated below 800 mT, or even 1 T (Fig. 3C, E), suggesting the co-existence of both low and high coercivity

minerals (Tauxe et al., 1996). The thermomagnetic curves reveal a Curie temperature of ~580 °C (Fig. 3F) or ~680 °C (Fig. 3D), suggesting that magnetite or hematite are the main magnetic minerals in these samples.

Fig. 3

Orthogonal plots of the results of the demagnetization of representative samples are illustrated in Fig. 4. In the case of most of the reddish clay samples, two remanence components can be recognized. The first is removed at a temperature of 200-300 °C. After removal of this component, the directions of the second component become relatively stable and the remanence decays univectorially towards the origin until the unblocking temperature at 580-680 °C (Fig. 4A, B). This component is regarded as the ChRM and its direction can be identified mostly between 440 °C and 640 °C. However, in the case of some green-greyish and grey clay and silty clay samples, the directions of the magnetization are stable at low temperature and decay univectorially towards the origin with an unblocking temperature of 360-440 °C (Fig. 4C). In the case of the latter samples, the results indicate that ferrimagnetic minerals like greigite are the main magnetic carriers (Dunlop and Özdemir, 2001; Hu et al., 2002), which is consistent with the results of the magnetic hysteresis loops and thermomagnetic curves (Fig. 3A, B). This low temperature remanence is also interpreted as the ChRM and its direction can be identified between 240 °C and 400 °C.

The ChRM directions were determined using at least four demagnetization steps that define a linear trend toward the origin of orthogonal vector demagnetization plots.

Samples with a maximum angular deviation (MAD) more than 15° were excluded. Parallel samples were measured if the remanence was unstable and did not pass the confidence test. Because of horizontal rotation of the core during drilling, the declination data are unusable and therefore the construction of the magnetic polarity sequence of core WEDP05 is based solely on inclination values. 167 specimens with unstable remanent magnetizations were eliminated and 330 specimens with reliable ChRM directions were selected to establish the magnetic polarity sequence. The polarity chrons were defined by at least three successive data points. Several individual samples, although stably magnetized, had polarities which were inconsistent with those of adjacent samples (Fig. 5, represented by open circles), and were deemed to be unreliable for polarity determination.

Fig. 4

The magnetostratigraphy of core WEDP05 consists of three normal (N1-N3) and three reversed (R1-R3) polarity zones (Fig. 5). The green-greyish clay layer at the top of Unit 2 and within N1 is widely distributed within the Hetao Basin and Jilantai area, and has been dated to the late Pleistocene (Li et al., 2005; Chen et al., 2008; Jiang et al., 2013; Chen et al., 2014; Li et al., 2015). In addition, the ESR dating results indicate that the ages of the samples above 142 m (n=5) are less than 0.78 Ma. Compared with the geomagnetic polarity timescale (GPTS) (Gradstein et al., 2012), we infer that the long interval containing N1 and N2 is the Brunhes Chron. There are 24 data points with inconsistent reversed polarities during the Brunhes Chron (Fig. 5). R1, defined by five successive points, may be correlated to an indefinable excursion,

and the other non-consecutive data points may be interpreted in two ways: 1) they record a genuine geomagnetic excursion, or 2) are the result of core disturbance during drilling. These inconsistent data points have little effect on polarity assignment because they constitute only a small proportion of all of the data points within the Brunhes Chron. The lower reversed polarity sequence comprising R2 and R3 is assigned to the Matuyama Chron. Thus, the Brunhes/Matuyama (B/M) boundary (0.78 Ma) is at the depth of 143.41 m. Noteworthy are three samples below the depth of the proposed B/M boundary with the ESR ages younger than 0.78 Ma, which suggest that the latter are underestimated. Based on these correlations, the short normal polarity zone N3, which is defined by four successive data points, should correspond to the Jaramillo event. According to the correlation, the polarity boundaries within the section were obtained. The age of the top of core WEDP05 is assumed to be modern because the chronology of several profiles and drill cores not far from the location of core WEDP05 indicate a Holocene age for the uppermost sediments (J. Sun et al., 2006; Fan et al., 2013; Liu et al., 2014). The age of the bottom of the core is estimated to be approximately 1.68 Ma based on extrapolation of the average sediment accumulation rate in the Brunhes Chron. A time series was obtained by linear interpolation between the two boundary ages, and from this the major lithological boundaries are dated to approximately 0.07 Ma (Unit 1/2), 0.28 Ma (Unit 2/3), 0.47 Ma (Unit 3/4), 0.78 Ma (Unit 4/5) and 1.17 Ma (Unit 5/6).

Fig. 5

4.4. Sedimentary facies and paleoenvironmental results

4.4.1. Sedimentary facies results

Several sedimentary facies can be recognized based on visual observation. For example, the sediments of Units 2-3 are dominated by grey, and green-greyish clay and silty clay with distinct horizontal bedding in some layers (Fig. 2); and in addition, abundant ostracod shells are present. These characteristics suggest that the sediments were mainly deposited in a lacustrine environment. A similar lithology, characterized by a green-greyish clay layer, also occurs in Unit 4 and again indicates a lacustrine origin. Some of the clay and silty clay layers in Units 5-6 contain ostracods (Fig. 2E1), which suggests they were mainly deposited in a lacustrine environment.

With regard to the microtexture of the sediment grains, the aeolian sand grains is characterized by well-rounded shapes with typical dish-shaped depressions, whereas that of fluvial sand grains is characterized by sub-rounded shapes with typical V-shaped depressions. The lacustrine deposits exhibit a cleavage surface with striations, and loess grains are irregularly shaped and have dish-shaped depressions (Xie, 1984). According to these characteristics, the sample from 3.10 m (Fig. 6A) is dominated by fluvial sand, the samples from 93.37 m (Fig. 6B) and 243.90 m (Fig. 6D) are aeolian sand, and the sample from 179.08 m (Fig. 6C) is dominated by lacustrine material.

Fig. 6

In core WEDP05, typical fluvial deposits consist of a dominant fluvial saltation component (fluvial sand), a fluvial suspension component (fluvial silt), and an ultra-fine component (Fig. 7A). Typical aeolian sand exhibits a tri-modal grain-size

distribution consisting of a dominant aeolian saltation or rolling component (aeolian sand) with a modal size of around 120 μm , a minor short-term suspension (coarse aeolian dust) component and a long-term suspension (fine aeolian dust) component (Fig. 7B, F). The fluvial sand component has a modal size of around 270 μm and is larger than that of the aeolian sand component (Fig. 7A). Typical lacustrine deposits are characterized by a bimodal grain-size distribution consisting of a dominant suspended component (lacustrine clay) together with an ultra-fine chemogenic component; in addition, they have a minimum modal size of 10-20 μm (Fig. 7C). Likewise, some typical lacustrine sediments have a multi-modal grain-size distribution consisting of a dominant lacustrine clay component and other components including an ultra-fine chemogenic component, aeolian sand, fluvial sand and dust (Fig. 7D and E). In addition, the lacustrine deposits contain more ultra-fine components than the aeolian and fluvial deposits (Fig. 7).

Fig. 7

The modal size-frequency curves clearly indicate that five groups of grain-size components can be identified, with boundaries of 3, 25, 100, and 252 μm (Fig. 8A). Based on comparison with the typical sediment samples of known origin, the origins of these groups can be determined. The <3 μm group is likely to be of pedogenic or chemogenic origin (Sun et al., 2002; 2011b). The 25-100 μm group with a similar modal size range to the coarse component in loess sediments (Sun et al., 2002; 2011b) corresponds to the dust in short-term suspension (coarse dust) (Fig. 8A, B). The group between 100-252 μm is defined as an aeolian sand component (Fig. 8A, B), since the

size range is typical of the desert sand in the Mu Us and Badain Jaran Deserts (Sun et al., 2002; Wang et al., 2015). The $>252\ \mu\text{m}$ group, with the modal size of $\sim 350\ \mu\text{m}$, falls within the typical size range of the fluvial saltation component (Sun et al., 2002). Throughout core WEDP05, typical fluvial sediments dominated by fluvial sand with mean size greater than $250\ \mu\text{m}$ (Fig. 7A) mainly occur in the top 5.5 m (Fig. 2). In the lacustrine deposits, the fluvial sand component generally has a low proportion (Fig. 7D). Therefore, the group $>252\ \mu\text{m}$ is defined as fluvial sand (Fig. 8A, B).

The group from $3\text{--}25\ \mu\text{m}$ is a combination of fine suspension components. According to the grain-size distribution of typical sediments (Fig. 7), it should contain long-term suspension dust (fine aeolian dust), a fluvial suspension component (fluvial clay) and a lacustrine suspension component (lacustrine clay). However, these components cannot be distinguished from the size-frequency curve (Fig. 8A) due to their having the same size range. However, in the proportion-frequency plot (Fig. 8C), a critical frequency value corresponds to the proportion of 50%, with frequency rising continuously above and below this boundary. This 50% inflection point is related to the main increase of the $3\text{--}25\ \mu\text{m}$ components. According to the lithological characteristics and the fitting results of typical sediments, most of the lacustrine deposits in core WEDP05 consist of a high proportion (general more than 50%) of the lacustrine clay (Fig. 7C, D, E), whereas the proportions of the fine aeolian dust and fluvial clay in all types of sediments are generally less than 50% (Fig. 7). Therefore, the proportion with more than 50% of components with a modal size from $3\text{--}25\ \mu\text{m}$ represents lacustrine clay, thereby indicating a dominantly lacustrine sedimentary

environment. Due to their having a similar proportion range, the type of component with a proportion below 50% is determined by the concomitant dominant component, because it is the attachment to the dominant component in the same sedimentary system. Therefore, the component accompanying fluvial sand is fluvial clay, while that accompanying aeolian sand or coarse dust is fine dust. The origins of each component in the WEDP05 core are clearly illustrated in the size-proportion plot (Fig. 8B).

Fig. 8

The long-term variation of the modal sizes and proportions of the main component could, to some extent, provide some evidence for identification of the sedimentary facies. The variations of the proportion and modal size of the main components are illustrated in Fig. 9. Accordingly, the sediments of Unit 1 are interpreted as mainly fluvial and aeolian sand. The fluvial component dominates the uppermost 5.5 m of this unit. The aeolian sand dominates most of deposits between 5.5 and 11.3 m, whereas a fluvial component dominates in some interlayers. The sediments of Units 2-3 are interpreted as fluvio-lacustrine clay mixed with aeolian dust. The proportions of the lacustrine and aeolian dust components increase significantly and alternately dominate the sediments. A fluvial component with a large modal size commonly accompanies the lacustrine component. The sediments of Unit 4 are interpreted as aeolian sand interbedded with fluvio-lacustrine deposits. The aeolian sand component dominates the sand layers from 109.4-87.5 m and from 141.6-125.5 m, and the fluvial and lacustrine components are almost absent in these

sand layers. The lacustrine component dominates the clay and silty clay layers from 125.5-109.4 m. The sediments of Units 5-6 are interpreted as fluvio-lacustrine mixed with aeolian dust interbedded with aeolian sand. In the clay and silty clay layers from 182.8-160.4 m and 235.5-207.2 m, the lacustrine and fluvial content increases significantly, whereas the content of aeolian sand is substantially reduced. In contrast, in the sand layers from 207.2-182.8 m and 269.4-235.5 m, the content and mean grain-size of the aeolian sand are significantly increased, whereas lacustrine and fluvial material only occurs in narrow layers.

In summary, based on the combination of visual observation of lithology, and the analysis of grain morphology and grain-size, the sediments of core WEDP05 exhibit the characteristics of fluvio-lacustrine deposits mixed with aeolian dust, alternating with aeolian sand. From 235.5-207.2 m (L1), 182.8-160.4 m (L2), 125.5-109.4 m (L3) and 87.5-11.3 m (L4), the sediments consist primarily of fluvio-lacustrine deposits mixed with aeolian dust. From 269.4-235.5 m, 207.2-182.8 m, 141.6-125.5 m, 109.4-87.5 m and 11.3-5.5 m, the sediments consist primarily of aeolian sand. The uppermost 5.5 m of the core consists of fluvial sand.

Fig. 9

4.4.2. Sediment color

Color is a visual property of sediment and is widely used in paleoenvironmental and paleoclimatic reconstruction (e.g. Balsam et al., 1999; Helmke et al., 2002; Sun et al., 2011a; Wang et al., 2014; Wang et al., 2015). In general, sediment deposited in a well-drained environment with a red color (high a^* value) is considered to result from

the presence of hematite (α -Fe₂O₃), which is formed in a warm and dry climate; while a yellowish color (low a^* value) is ascribed to goethite (α -FeOOH), which is formed in a colder and more humid environment (Cornell and Schwertmann, 2003; Viscarra Rossel et al., 2006; Wang et al., 2014). Within the drill core, a^* values of the aeolian sediments are general low (less than 4) (Fig. 9), and such values indicate that the sediments were deposited in a relatively cold environment with limited hematite formation, and may thus indicate local or regional climatic cooling. In lacustrine strata, color can be used as a rough indicator of water depth (Wu and Li, 2004; Song et al., 2005). In deep lakes, hematite is more easily reduced to goethite and this mineralogical transformation tends to impart a greenish color to the sediments, compared to the sediments of shallow lakes. Thus, the sediments of deep lakes typically have lower a^* values than those of shallow lakes. In core WEDP05, the lower a^* values of the lacustrine sediments in L3 and L4 indicate that the lake water experienced relatively strongly reducing conditions compared to those corresponding to the depths of L1 and L2 (Fig. 9). Thus, the extent and depth of the paleolake during L3 and L4 may have been greater than during L1 and L2.

5. Discussion

5.1. Paleoenvironmental evolution of the Hetao Basin

The results from core DR01 enable the Cenozoic paleoenvironmental history of the Hetao Basin to be reconstructed. Geological survey results have revealed that the Palaeocene and Eocene sedimentary rocks in the Hetao Basin are absent, resulting in an unconformity between the Cretaceous and Cenozoic sedimentary sequence (CEA,

1988; Zhang et al., 1983; Cai et al., 1988). This sedimentary hiatus was found in core DR01 between the early Cretaceous and Pliocene strata (Xu et al., 2012; Fig. 2), which suggests that the Hetao Basin may originally have had a relatively high relief and was therefore significantly affected by erosion from the Mesozoic to the early Pliocene. The more than 2000 m thickness of sediment in core DR01 representing post-Miocene time suggests that the area subsided to form a sedimentary basin and began to accumulate detrital sediment. During the Pliocene, the lithology of core DR01 core mainly consists of light-brown clay, silty clay (Fig. 2), which was defined as fluvio-lacustrine sediments (CEA, 1988; Zhang et al., 1983), suggesting that a fluvio-lacustrine environment developed in the basin. During the Quaternary, geological survey revealed that a fluvio-lacustrine environment frequently developed in the basin (CEA, 1988). In core DR01, the presence of grey and green-greyish clay and silty clay (Fig. 2) also suggest that a lacustrine environment existed in the basin. On the other hand, some fine sand grains in the Quaternary strata of the basin, as well as the fine sand sediments in core DR01, are well-rounded and sorted with a loose structure (Zhang et al., 1983; Deng et al., 2007; Xu et al., 2012), possibly indicating an aeolian origin. More importantly, sedimentary facies results indicate an aeolian origin of most of the sand layers in core WEDP05. This evidence suggests that a desert environment may have occasionally occurred in the basin. Due to the low recovery rate of core DR01, there are several differences in the petrographic description between this core and core WEDP05. Nevertheless, geophysical logging results for core DR01 indicate large fluctuations in resistivity (from 0~45 $\Omega\cdot\text{m}$) in

the Quaternary strata (Xu et al., 2012), suggesting a similar pattern of alternating layers of clay, silty clay and sand as occurs in core WEDP05, and thus indicating a similar paleoenvironment. The fluvio-lacustrine sediments intercalated with aeolian sand sediments suggest that alternations between a fluvio-lacustrine and aeolian environment occurred throughout the entire Quaternary. This scenario may reflect the evolution of the Yellow River in response to tectonic activity or to periodic climatic fluctuations.

Based on the variation of the sedimentary facies, grain size and redness (a^*) values, together with the time scale obtained from core WEDP05, the paleoenvironment of the Hetao Basin since the early Pleistocene can be divided into four stages (Fig. 10).

Fig. 10

Stage 1 (~1.68 Ma to ~0.68 Ma): The lithology of this stage is characterized by alternating layers of lacustrine clay and silty clay and aeolian sand (Fig. 9). The lacustrine strata indicate at least two intervals during which a paleolake occupied the Hetao Basin: L1, from ~1.47 - ~1.30 Ma; and L2, from ~1.17 - ~1.07 Ma (Fig. 10). Grain-size analysis results indicate that the fluvial components and lacustrine component coexisted and dominated the strata, suggesting the lake may have been fed by a river, especially after the formation of the Yellow River which flowed through the Hetao Basin (Fan et al., 2010). The high color a^* values, together with the relatively coarse mean grain-size of these two lacustrine layers (Fig. 10), indicate a shallow lake and an oxidizing sedimentary environment. Investigations of the regional

geology of the Hetao Basin have revealed that the early Pleistocene deposits are mainly composed of reddish clay and silt clay and yellowish sand (CEA, 1988). These characteristics are consistent with the lithology of stage 1 in core WEDP05, suggesting that a lake environment may have occupied a large part of the basin. During the intervals of lake regression, from ~1.30 - ~1.17 Ma and from ~0.85 - ~0.68 Ma, the strata were dominated by aeolian sand (Fig. 10), which may have been mobilized and transported by the winter monsoon from areas of exposed lake sediments to downwind areas, such as the Hobq Desert (Fig. 1B), to form the sand dunes. From the foregoing, the history of the Hobq Desert can be extended back to the early Pleistocene.

Stage 2 (~0.68 Ma to ~0.47 Ma): The reddish lacustrine clay interbedded with greyish-green lacustrine clay layer (Fig. 9) suggests that a lake environment recurred in the basin from ~0.68 Ma to ~0.60 Ma (Fig. 10). This stage corresponds to the first appearance of greyish-green lacustrine strata in the drill core (Fig. 2). The lower a^* color values of the strata, compared to that of the L1 and L2 intervals, indicate a more strongly reducing sedimentary environment. This aspect of the lithology, together with the horizontal bedding of the strata, further suggests a deeper lake environment in the Hetao Basin. However, in core ZKHB, drilled to a depth of 250.03 m in the Hohhot Depression (Fig. 1C), a sedimentary hiatus occurred from about 0.8 - 0.5 Ma (J. Li et al., 2007; Fig. 11). In addition, in the Haojiayao (HJY) core, drilled to a depth of 118.45 m in the margin of the Hohhot Depression (Fig. 1C), reddish clay deposited prior to 476 ka was classified as loess-like sediment (Jiang et al., 2013; Fig. 11). This

evidence suggests that the lake did not occupy the entire basin, and that some parts remained a sub-aerial environment (e.g the Hohhot Depression). After this lacustrine stage, the deposition of greyish and yellowish aeolian sand layers indicates a return to a desert environment within the basin from ~0.60 Ma to ~0.47 Ma (Fig. 10).

Stage 3 (~0.47 Ma to the last interglacial): The green-greyish and grey clay and silt clay, together with the lowest values of a^* color within the entire WEDP05 drill core (Fig. 9), suggest the formation of the deepest lake environment since the early Pleistocene (Fig. 10). Similar lacustrine strata also occur in drill core DR01, with a thickness of almost 90 m (Fig. 11). Furthermore, in drill core ZKHB (Fig. 1C), typical lacustrine clay and silty clay layers are present which are dated to ~363 - ~86 ka (J. Li et al., 2007; Fig. 11). In drill core HJY (Fig. 1C), lacustrine strata with a thickness of over 45 m were deposited after ~293 ka (Jiang et al., 2013; Fig. 11). This synchronous development of lacustrine deposits suggests that most of the Hetao Basin was occupied by a lake during the middle-late Pleistocene. In fact the lake water may have extended to the southern part of the Hobq Desert and to the eastern part of the Tuoketuo (Fig. 1C). During the last interglacial, lake deposits also can be recognized in a 120.5-m-long core WL12ZK-1 from the southern Ulan Buh Desert (Fig. 1C), with an age range from ~155 ka to ~87 ka (Li et al., 2015; Fig. 11); and in sections from the Tuoketuo area (Fig. 1C) with an age range from ~122 ka to ~82 ka (Li et al., 2005) or from ~150 ka to ~100 ka (Jiang et al., 2013). This evidence suggests that the lake expanded to the western part of the Jilantai area and Ulan Buh Desert (Fig. 1C); the lake has been named the “Jilantai-Hetao Megalake”, and with an estimated surface

area of $3.4 \times 10^4 \text{ km}^2$, it was the largest confirmed inland lake in China during the late Quaternary (Chen et al., 2008).

Fig. 11

Stage 4 (last glacial and Holocene): During the last glacial, the lacustrine sediments disappeared and aeolian sands dominate the strata. The significant increase of the content and grain-size of the aeolian sand component may indicate the occurrence of large aeolian sand dunes in the Hetao Basin and Hobq Desert. Studies of aeolian deposits suggested that deserts in southern Inner Mongolia expanded and the stabilized dunes were reactivated at the same time (Lu et al., 2005; J. Sun et al., 2006; Fan et al., 2013; Chen et al., 2014; Wang et al., 2015; Li et al., 2015), suggesting that a very arid climate dominated southern Inner Mongolia during the last glacial. During the Holocene, typical fluvial sand sediments, with the highest grain size within the drill core, dominate the strata and suggest that a fluvial-alluvial environment developed within the basin. At the same time, wetland or lakes appeared in the Ulan Buh Desert (Chen et al., 2014; Li et al., 2015), Badain Jaran Desert (e.g. Yang et al., 2010, 2012; Wang et al., 2015), and Tengger Desert (Madsen et al., 2003; H. Zhang et al., 2004; Long et al., 2012). In addition, all of the Chinese sand fields were almost completely covered by vegetation during the early Holocene (Lu et al., 2013). All of this evidence suggests that southern Inner Mongolia became more humid during the Holocene.

5.2. Formation of the Hobq Desert

As one of several proximal deserts, the Hobq Desert is located on the windward

side of the Loess Plateau (Fig. 1B) and is an important potential sediment source for the Loess Plateau (Sun, 2002; Rao et al., 2008). Its formation and evolution history may improve our understanding of the linkage between sediment source and the origin of the Loess Plateau. At the base of the WEDP05 drill core, the yellowish sand layer, with a thickness of more than 30 m, has a mean grain-size ranging from 110-170 μm (Figs. 2 and 9). The sand grains have well-rounded shapes with typical dish-shaped depressions (Fig. 6D), and are defined as aeolian sands. The grain size distributions are typical of aeolian sands (Fig. 7F) and the grain-size analysis results suggest that this sand layer is dominated by an aeolian sand component (Fig. 9). Both of these characteristics suggest that this sand layer is aeolian. The presence of this layer may indicate the presence of sand dunes in the Hobq Desert, suggesting that the desert formed in the early Pleistocene, at least since ~ 1.65 Ma.

Chronological work on the aeolian sand overlaying palaeosols or lacustrine layers suggests that the modern desert landscape of the Hobq Desert was formed in the late Pleistocene (S. Li et al., 2007; Fan et al., 2013) or Holocene (J. Sun et al., 2006). However, Yan (2004) suggest that the formation of the Hobq Desert can be traced to the middle Pleistocene since the sand dunes had already formed at that time in spite of the subsequent development of palaeosols developed during the early Holocene. Recently, Yang et al. (2016) suggested a Holocene age for the sand sea of the Hobq Desert based on optically stimulated luminescence (OSL) ages of the aeolian sand overlying bedrock. Because of the sedimentary hiatus between the sand dunes and the underlying bedrock, the sedimentary sequence from Yang et al. (2016) may not be

sufficient to reflect the complete evolution history of the Hobq Desert, and the age of the sand layers may represent the activity of the sand dunes and expansion of desert landscapes during the early Holocene, which is consistent with our interpretation that the large aeolian sand dunes formed in the Hobq Desert since the last glacial.

With regard to the formation ages of proximal deserts to the Loess Plateau, recent chronological work indicates that the formation of the Badain Jaran Desert occurred at ~1.1 Ma (Wang et al., 2015) and the Tengger Desert at ~0.9 Ma (Z. Li et al., 2014), indicating a much younger formation age than that of the Hobq Desert. Several models have been suggested for the origin of the sediment sources of deserts in northwest China based on different geological and geomorphological settings: 1) denudation of surrounding mountains; 2) weathering of the underlying parent rocks; 3) reworking of fluvial-lacustrine deposits; and 4) desertification (Zhu et al., 1980). In the case of the Badain Jaran and Tengger Deserts, the desert sands were found to be transported from upwind alluvial/fluvial fan areas by the Asian winter monsoon and the Westerlies (Z. Li et al., 2014; Wang et al., 2015; Hu and Yang, 2016). In the case of the Hobq Desert, several lines of evidence indicate that the aeolian sand mainly originated from local fluvial and lacustrine sediments transported by winds associated with the near-surface Asian winter monsoon: 1) Aeolian sand and fluvial-lacustrine sediments commonly co-exist in the strata of core WEDP05 (Fig. 2); 2) active sand dunes are common around river banks and on the margins of lakes; and 3) these eroding river banks and lake margins are located upwind of the Hobq Desert, under the pathway of the Asian winter monsoon (Fig. 1), and especially during the cool and

dry season, they are subject to erosion and the sediments are transported to the Hobq Desert, and even to the Loess Plateau.

Although the mode of sediment transport to the Hobq Desert is similar to that of the Badain Jaran and Tengger Deserts, its sediment sources are different. Previous work suggested that alluvial/fluvial fan areas of the Heihe River and Shiyanghe River are the dominant source for sandy material in the Badain Jaran and Tengger Deserts (Yang et al., 2012; Z. Li et al., 2014; Wang et al., 2015; Hu and Yang, 2016). Thus, the formation of both deserts was to a large extent determined by the evolution of the Heihe River and Shiyanghe River, the formation ages of which were both dated to 1.2-0.9 Ma (Pan et al., 2003, 2007, 2016). In the case of the Hobq Desert, the aeolian sand mainly originated from the reworking of the fluvial-lacustrine sediments in the Hetao Basin. These deposits may be mainly carried into the basin by the Yellow River, following the formation of its upper reaches, and which flowed into the Hetao Basin; in addition, it is noteworthy that the Yellow River is an abundant source of clastic material. Therefore, it can be concluded that the formation and evolution of the Yellow River has determined the development of the Hobq Desert. For this reason, the age of the Hobq Desert can be traced to the early Pleistocene, since the formation of the upper reaches of the Yellow River and its drainage within the Hetao Basin occurred at least since ~1.6 Ma (see Discussion 5.3). It is therefore clear that the Hobq Desert is older than the Badain Jaran and Tengger Deserts.

In addition, work on the origin of the Loess Plateau has revealed that a genetic link exists between the Yellow River and associated fluvial systems and the formation

of the loess deposits on the Loess Plateau (Stevens et al., 2013; Bird et al., 2015; Shang et al., 2016; Licht et al., 2016). Additionally, a recent study revealed that the sediments of the western Mu Us Desert and the Loess Plateau originate from Yellow River alluvium that was deposited in the Yinchuan-Hetao Graben, and was subsequently redistributed by winter monsoon winds after 3.6 Ma (Nie et al., 2015). As explained above, sand dunes in the Hobq Desert also originate from the Yellow River sediments in the Hetao Basin, and the Hobq Desert and Hetao Basin are located directly upwind of the Loess Plateau under the pathway of the dust transported by the East Asian winter monsoon circulation (Fig. 1B). Thus, during the period of sand dune formation in the Hobq Desert, a certain proportion of finer-grained materials are likely to have been transported by wind to the Loess Plateau. From the foregoing, the significantly older age of the Hobq Desert, compared to the Badain Jaran and Tengger Deserts, may provide evidence to support the important role of the Yellow River in the formation of the Loess Plateau (Stevens et al., 2013; Bird et al., 2015; Nie et al., 2015; Shang et al., 2016; Licht et al., 2016).

With regard to the formation of a desert environment, some authors suggest that an active aeolian system may be sustained under relatively humid conditions in the case of sufficient sediment supply and the availability of sediment for aeolian transport as a result of, e.g., fluvial activity (Bullard and Livingstone, 2002; Cohen et al., 2010; Al-Masrahy et al., 2015; Halfen et al., 2015; Nottebaum et al., 2015). In this context, climate exerts a fundamental control on the expansion and contraction of sand dune-field margins (e.g., Herries, 1993; Clarke and Rendell, 1998; Al-Masrahy

et al., 2015). In the case of the Hobq Desert, the activity of the Yellow River ensured sufficient sediment supply for sand dune formation (Fig. 1B). Therefore, the enhanced sediment supply of the Yellow River may have been a fundamental factor for the formation of sand dunes in the Hobq Desert. On the other hand, the alternating deposition of fluvial-lacustrine and aeolian sand layers in the sedimentary sequence of core WEDP05 suggests that the sand dunes were formed only after the desiccation of paleolakes, which made previously-deposited fluvial-lacustrine sediments available for transport and thus providing plentiful material for the formation of the desert dunes of the Hobq. Thus, the absence of lake systems in the basin may have been another crucial factor for the development of sand dunes in the Hobq Desert.

However, it is difficult to believe that the desert environment of the Hobq Desert formed under a humid climate. The aeolian sand accumulations in core WEDP05 have a duration of several tens of thousands of years (Fig. 10), suggesting that a relatively long-term dune environment was sustained in the Hobq Desert. This pattern would require not only sufficient sediment supply but also long-term sand availability and mobility. Such a scenario is almost impossible under a long-term humid climate because the availability and mobility of sand material is limited due to increased vegetation cover, high effective moisture and low wind transport capacity (Hesse et al., 2006; Mason et al., 2011). In contrast, in an arid climatic phase, increased sand availability, mobility and deposition are promoted by a high wind strength, reduced vegetation cover and low effective moisture (Wiggs et al., 1995; Lancaster et al., 1998; Hesse et al., 2006; Mason et al., 2011). This is the main reason that most of the

modern desert dunes in China are distributed in arid and semiarid areas (Fig. 1A). In conclusion, we suggest that an arid climate probably played an important role in the formation of sand dunes in the Hobq Desert.

5.3. Formation and evolution of the Yellow River

The formation of a series of narrow basins, distributed in a rounded rectangular shape around the Ordos Plateau (Fig. 1), is the key to understanding the evolution of the upper and middle reaches of the Yellow River. Among this series of basins, the evolution of the Hetao Basin played a crucial role in the formation of the Yellow River. The Yellow River drainage around the Ordos Plateau is generally considered to have formed as a result of the connecting of a series of isolated paleorivers and/or paleolakes in basins around the Hetao Basin (e.g. Li, 1988; Jiang et al., 2007; Pan et al., 2012). The water source of the paleolake formed within the Hetao Basin is closely related to the Yellow River after its formation. Thus, the paleoenvironmental evolution of the Hetao Basin may provide important evidence for the evolution history of the Yellow River.

From the Miocene to the Quaternary, over 4100 m of fluvial-lacustrine sediments accumulated in the Hetao Basin, accompanied by the progressive subsidence of the basin, which suggests that the Hetao Basin was a depocenter, at least since the Miocene (CEA, 1988). In core WEDP05, the uppermost 5.5 m of sediments accumulated in a normal Yellow River environment, characterized by coarse sand with the largest grain size found within the entire sequence (Fig. 9). The underlying strata, which include lacustrine clay and silty clay intermixed with fluvial components

(Fig. 9), suggest an even more extensive fluvio-lacustrine environment. Although this type of sediment is dated only to ~1.68 Ma in core WEDP05, similar sediment types can be traced back to the Pliocene in core DR01 (Xu et al., 2012; Fig. 2). This suggests that the fluvio-lacustrine environment in the Hetao Basin can be traced back to the Pliocene. Similarly, Pliocene-Quaternary fluvio-lacustrine sediments are also distributed in other fault basins, including the Yinchuan and Fenwei Basins, along the current courses of the Yellow River (CEA, 1988; Lin et al., 2001; Fig. 1B). These observations suggest that all of these basins around the Ordos Block may have formed as early as the Pliocene, and perhaps were proto-river channels.

However, this does not mean that the Yellow River had formed by that time, because no evidence is available which supports the interconnection of these basins via the Yellow River during the Pliocene. Recent work suggested a Pliocene age of the Yellow River based on the similarity of the provenance of Pliocene conglomerates deposited in the Linxia and Lanzhou Basins (J. Li et al., 1996, 2014; Sun et al., 2011c) and typical Yellow River terraces (Nie et al., 2015). However, the similar provenance of these two sediment types may not be an indicator of the formation of the Yellow River, because they both originated from the NE Tibetan Plateau irrespective of the formation of the Yellow River. Systematic geomorphological work conducted in the Lanzhou and Zhongwei area (Fig. 1B) suggests that the upper reaches of the Yellow River from Yinchuan Basin formed from ~1.7 - ~1.6 Ma (Li, 1991; Li et al., 1996, 1997; Xing et al., 2002). Because of the geographical connectivity of the Hetao and Yinchuan Basins (CEA, 1988), the Yellow River may have flowed across the

Yinchuan Basin and discharged into the Hetao Basin at least since that time. Therefore, we suggest that the Yellow River may have entered the Hetao Basin and formed its upper reaches at least since ~1.6 Ma. In the downstream areas of the Hetao Basin, geomorphological work has revealed that the fluvial terrace of the Yellow River formed simultaneously at ~1.2 Ma in the Jinshan Gorge (Cheng et al., 2002; Pan et al., 2012; Hu et al., 2016), in the Fenwei Basin (Yue et al., 1997; Sun, 2005; Hu et al., 2012), in the Sanmen Gorge and in the Kouma area (Pan et al., 2005; Su et al., 2009; Kong et al., 2014). This indicates that the Yellow River flowed through the Jinshan Gorge, and was connected with the Fenwei Basin, the Sanmen Gorge and the Kouma area at least since ~1.2 Ma. All of this evidence suggests that the Yellow River flowed out of the Hetao Basin and had completely established its current drainage pattern by at least ~1.2 Ma.

After the formation of the Yellow River by ~1.2 Ma, the paleoenvironment, especially the lake environment of the Hetao Basin, was closely related to the evolution of the Yellow River. The following factors may have been significant: (1) A tectonically-driven increase in the relative altitude of the basin outlet, resulting from the subsidence of the basin and/or the uplift of Jinshan Gorge, generating a tectonic barrier, could have caused the formation of the lake in the basin. (2) An increase in the discharge of the Yellow River could have promoted the formation of a deep lake under the conditions described in (1), above. (3) Outlet erosion and/or sediment overfilling by the Yellow River could have resulted in the disappearance of the lake. The development of numerous Yellow River terraces in the area downstream of the

Hetao Basin since ~1.2 Ma (Yue et al., 1997; Pan et al., 2005; Sun, 2005; Su et al., 2009; Hu et al., 2012, 2016; Kong et al., 2014; Meng et al., 2015) may suggest the presence of an open drainage system in the basin after the formation of the Yellow River. Although a humid climate could have increased the discharge of the Yellow River, it could not have caused the formation of a lake in such an open basin system and therefore we suggest that tectonic activity was the dominant factor driving the formation of the lake in the Hetao Basin.

Fig. 12

Prior to ~0.47 Ma, the channel of the Yellow River may have widened thereby inducing lake formation within the basin from ~1.17 - ~1.07 Ma and from ~0.68 - ~0.60 Ma. The very thick Quaternary sediment sequences deposited within the graben basins around the Ordos Block suggest that they were subjected to continuous subsidence during the Quaternary (CEA, 1988; H. Chen et al., 2015; Shi et al., 2015). This process was regarded as a response to the progressive northeastward propagation of the Tibetan Plateau (CEA, 1988; Zhang et al., 1998; H. Chen et al., 2015; X. Chen et al., 2015; Shi et al., 2015), which also caused the uplift of the Jinshan Gorge and the formation of the Yellow River terrace (Cheng et al., 2002; Hu et al., 2016). A relatively high basin outlet resulting from this differential tectonic movement could have widened the Yellow River channel in the basin to form a lake. This scenario is supported by the synchronous development of a lake environment in the Hetao Basin and by the Yellow River terraces in the Jinshan Gorge (Hu et al., 2016; Fig. 12A). Basin subsidence may have been the dominant factor for the development of various

lakes in the graben basins around the Ordos Block, especially in the case of several basins in Shanxi Graben, which are not located in the current course of the Yellow River (Fig. 1B). In addition, the different depth of the lakes formed within the basin during this period may have been driven by a climatic factor, and the greater extent of the lake during the L2 interval compared with that during the L3 interval may have resulted from the enhanced discharge of the Yellow River.

From ~0.47 Ma to the last interglacial, a widened Yellow River channel may have caused the formation of a deeper lake within the Hetao Basin. Several lines of evidence suggest that this was a combination of tectonically-driven rapid basin subsidence and uplift of its outlet: 1) Analysis of the salt content in the lacustrine sediments indicates that the lake was saline (Jia et al., 2001), suggesting a long residence time of the lake water. 2) This period of deep lake formation in the Hetao Basin is contemporaneous with a period of accelerated incision of the Yellow River at the Jinshan Gorge (Hu et al., 2016; Fig. 12C), the latter suggesting strong tectonic activity which not only induced rapid uplift of the Jinshan Gorge but also caused the rapid subsidence of the basin. The rapid rising Jinshan Gorge may have acted as a barrier promoting the formation of the deep lake. 3) Stratigraphic and chronological studies of cores HJY and ZKHB have revealed that stratigraphic unconformities occurred at ~0.5 and ~0.1 Ma (J. Li et al., 2007; Jiang et al., 2013; Fig. 11), indicating the uplift of the easternmost part of the Hetao Basin (i.e. the outlet of the basin, Fig. 1C) at ~0.5 and ~0.1 Ma. These dates correspond to the ages of two major deep lakes developed at ~0.47 and ~0.13 Ma in the basin, suggesting a causal relationship

between the raised outlet of the basin and the formation of these deep lakes. In addition, a strengthening of the Asian summer monsoon which migrated significantly northwards after ~0.5 Ma (e.g. An et al., 1991; Xiao and An, 1999; An et al., 2005; Y. Sun et al., 2006; Sun et al., 2010), resulting in an increase in surface runoff in the Hetao area and in the discharge of the Yellow River, may also have made an important contribution.

During the last glacial and Holocene, the desiccation of the paleolake and the deposition of fluvial sediments within the Hetao Basin suggest that the Yellow River resumed incision of the Jinshan Gorge. This is also evidenced by the wide distribution of the highest Yellow River terrace formed during the late Pleistocene in the vicinity of the Jinshan Gorge (Qiu et al., 2008; Zhang et al., 2009; Guo et al., 2012; Fu et al., 2013; Liang et al., 2015). In addition, stratigraphic work has revealed that the final disappearance of the paleolake in the Sanmen Gorge and Fenwei Basin both occurred during the late Pleistocene (Wang et al., 2002; Z. Zhang et al., 2004; Jiang et al., 2007; Prins et al., 2009; Qiu et al., 2015), which is almost synchronous with the disappearance of the paleolake within the Hetao Basin (Li et al., 2005; Chen et al., 2008; Jiang et al., 2013; Chen et al., 2014; Li et al., 2015; Jia et al., 2016; and this study). The consistency in the disappearance of the paleolake along the course of the Yellow River during the late Pleistocene may have been the result of intense Yellow River erosion during the late Pleistocene due to the strong uplift of the Tibetan Plateau. This phase of tectonic activity, namely the Gonghe Movement (Li, 1991), caused both the integration of the Yellow River within the inner Tibetan Plateau (Li et al., 1991;

Perrineau et al., 2011) and the steepening of its gradient. Both of these events enhanced the erosive capacity of the Yellow River, which enabled it to cut through dams and enlarge the channel in the downstream areas, triggering the desiccation of the paleolakes along its course. On the other hand, the increased erosive capacity of the Yellow River would have resulted in the increased transport of clastic material transported from NE Tibetan Plateau into the downstream sedimentary basins, which may have been an important factor in the shrinking and desiccation of the paleolake, since infilling of the basin would have eliminated the space available for water storage. This process would not only have made an important contribution to the demise of the paleolake but would also have increased the supply of sand material to form dunes, and thus may have been an important reason for the formation of sand dunes after the disappearance of the lake in the basin.

6. Conclusions

The Hetao Basin was uplifted and eroded from the early Cenozoic until the Pliocene, and subsequent subsidence resulted in the accumulation of more than 2000 m of detrital material since the early Pliocene. A fluvio-lacustrine environment occurred in the basin during the Pliocene, and there was an alternation between fluvio-lacustrine and aeolian environments throughout the entire Quaternary. Since the early Pleistocene, at ~1.68 Ma, a fluvial lacustrine system again occupied the basin during the following intervals: ~1.47 - ~1.30 Ma, ~1.17 - ~1.07 Ma, ~0.68 - ~0.60 Ma and from ~0.47 Ma to the last interglacial. A desert landscape developed within the basin during the stages of lake regression: ~1.60 - ~1.47 Ma, ~1.30 - ~1.17

Ma, ~1.07 - ~0.68 Ma, ~0.60 - ~0.47 Ma and the last glacial.

The formation of the Hobq Desert can be traced back to the early Pleistocene. The sediments of the desert are derived from the erosion and transportation of exposed fluvio-lacustrine sediments within the Hetao Basin by near-surface winds associated with the Asian winter monsoon. The formation of sand dunes in the Hobq Desert was the consequence of sufficient sediment supply by the Yellow River or of the vanishing of the paleolake within the Hetao Basin; an arid climate may also have made an important contribution.

The channel of the Yellow River within the Hetao Basin may have formed since the Pliocene; however, the current drainage system of the Yellow River was completely established by at least ~1.2 Ma. The relative height of the basin outlet, resulting from the subsidence of the basin and/or the uplift of Jinshan Gorge may have been the primary driving force behind the formation of the lake environments in the Hetao Basin. The downcutting of the Yellow River and/or infilling of the basin may have caused the desiccation of the paleolake in the Hetao Basin and other basins along its course.

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Figure Captions

Fig. 1. (A) Distribution of desert (yellow) and loess (red) in northern China. Desert distribution data are from <http://westdc.westgis.ac.cn>, and the distribution of loess is from Liu (1985). The black dashed line indicates the limit of the Asian summer monsoon identified by Gao (1962). (B) The proximal desert of the Loess Plateau and the fault-subsiding belts on the periphery of the Ordos Block (modified from CAE, 1988). YC-Yinchuan Graben, HT-Hetao Graben, SX-Shanxi Graben, FW-Fenwei Graben. The upper (U), middle (M) and lower (L) reaches of the Yellow River are divided by white lines. (C) Geological map superimposed on the topography of the Hetao Basin, and the locations of the WEDP05, DR01 and other drill cores referenced in the text. The contours (m) represent the thickness of the Quaternary fluvio-lacustrine deposits within the basin and are from CAE (1988). JD-Jilantai Depression, DU-Dengkou Uplift, LD-Linhe Depression, XU-Xishanzui Uplift, BD-Baiyanhua Depression, BU-Baotou Uplift, HD-Hohhot Depression.

Fig. 2. Lithology of the DR01 and WEDP05 drill cores and photographs of typical sediments in core WEDP05. The ages of core DR01 were determined by regional stratigraphical correlations and as defined by Xu et al (2012). Black circles next to the core WEDP05 log represent the photographed locations. (A and A1) greyish and yellowish sand in Unit 1; (B and B1) grey clay and silty clay in Unit 2; (C and C1) green-greyish clay with conspicuous horizontal bedding in Unit 3; (D and D1) well-sorted greyish sand in Unit 4; (E and E1)

1390 reddish clay and silty clay in Unit 5 (abundant ostracod shells are present in
1391 the sediments); (F, F1, G and G1) well-sorted yellowish sand in Unit 6.

1392 Fig. 3. Magnetic hysteresis loops (left) and thermomagnetic curves (right) for
1393 representative samples from core WEDP05. The magnetic hysteresis loops are
1394 corrected for the paramagnetic contribution. In the thermomagnetic curves, red
1395 (blue) lines represent heating (cooling) curves.

1396 Fig. 4. Orthogonal demagnetization diagrams, remanence intensity and equal area
1397 projections of typical samples from core WEDP05. The black solid (open)
1398 circles in the orthogonal plots represent projections on the horizontal (vertical)
1399 plane. In the equal area projections, the red solid (blue open) circles are
1400 projections onto the lower (upper) hemisphere.

1401 Fig. 5. Magnetostratigraphy and results of ESR dating of core WEDP05. Open circles
1402 represent samples which have a stable remanent magnetization but whose
1403 polarities are inconsistent with those of adjacent samples. The standard
1404 geomagnetic polarity time scale (GPTS) is from Gradstein et al (2012).

1405 Fig. 6. Scanning electron micrographs of quartz grains from sediments in WEDP05
1406 core: (A) angular morphology and different grain sizes of quartz grains from
1407 the sand layer at 3.10 m; (B) well-sorted and almost rounded quartz grains
1408 from the sand layer at 93.37 m; (C) cleavage surface of quartz grains from the
1409 reddish clay at 179.08 m; (D) well-rounded quartz grains with dish-shaped
1410 concavities from the sand deposits at 243.90 m.

1411 Fig. 7. Grain-size distributions of typical samples from core WEDP05. FE- fitting

1412 function with error, AS-aeolian sand, FS - fluvial sand, FC - fluvial clay, SSD -
1413 short-term suspension dust component, LSD - long-term suspension dust
1414 component, LC - lacustrine component, P - pedogenic/chemogenic
1415 component.

1416 Fig. 8. Grain-size analysis results of samples from the WEDP05 drill core: (A)
1417 frequency distribution of component size; (B) size-proportion plot of all of the
1418 grain-size components; (C) frequency distribution of component proportions.

1419 Fig. 9. Paleoenvironmental records from core WEDP05. The blue zones are intervals
1420 of dominantly lacustrine sediments in the core.

1421 Fig. 10. Time series of paleoenvironmental proxy indices from core WEDP05. The
1422 blue zones represent lacustrine phases.

1423 Fig. 11. Lithological correlation of five drill cores within the Hetao Basin. The
1424 different chronological methods are indicated by different colors. The
1425 lithology and ages for core WL12ZK-1 are from Li et al. (2015); those for core
1426 ZKHB are from Li et al. (2007b); those for core HJY are from Jiang et al.
1427 (2013); and those for core DR01 are from Xu et al. (2012). The correlations
1428 are based on changes in lithology and ages.

1429 Fig. 12. (A) Comparison of ages for the development of the paleolake within the
1430 Hetao Basin and the formation of the Yellow River terraces in the upstream of
1431 the basin in the Zhongwei area and downstream of the basin in Jinshan Gorge.
1432 The data for the terraces in the Zhongwei area are from Xing et al. (2002), and
1433 those for the terraces in Jinshan Gorge are from Hu et al. (2016). (B)

1434 Reconstructed average incision rate of the Yellow River in the Zhongwei area,
1435 which increased from ca. 0.10 m/ka to ca. 0.41 m/ka since ~140 ka. (C)
1436 Reconstructed average incision rate of the Yellow River in the Jinshan Gorge.
1437 Increased incision of the Yellow River at a rate of ca. 0.24 m/ka occurred since
1438 ~0.4 Ma.

1439 **Table Captions**

1440 Table 1 Electron Spin Resonance (ESR) dates and associated parameters.

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Table 1 Electron Spin Resonance (ESR) dates and associated parameters

Sample	Depth (m)	Water content (%)	U (ppm)	Th (ppm)	K (%)	D _e (Gy)	Dose rate (Gy/ka)	Age (ka)
KBQ-10	75	18	2.42±0.20	8.35±0.80	2.62±.020	737±65	4.352	170±15
KBQ-13	86	16	2.25±0.20	7.66±.070	2.13±.020	719±57	3.790	190±15
KBQ-20	101	16	1.38±0.12	4.15±0.40	2.03±.020	679±56	2.783	244±20
KBQ-28	120	13	1.40±0.10	4.13±0.40	2.86±0.20	1271±114	2.860	444±40
KBQ-30	142	17	1.02±0.10	2.55±0.25	1.97±0.18	1332±116	2.327	572±50
KBQ-31	154	13	1.60±0.10	4.96±0.45	2.45±0.20	1842±173	2.884	639±60
KBQ-33	169	12	1.85±0.15	6.66±0.60	2.17±0.20	2439±244	3.484	700±70
KBQ-35	240	15	1.42±0.14	4.53±0.45	1.79±0.15	3320±270	2.678	1240±100

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Fig. 12

Site location of drill core WEDP05

Magnetostratigraphy of core WEDP05 and the aeolian sand sediment at the base of the core