1 Drainage evolution of the Heihe River in western Hexi Corridor, China,
2 derived from sedimentary and magnetostratigraphic results

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6 **Abstract** The uplifting of the Tibetan Plateau has significantly changed the environment
7 in surrounding areas by delivering abundant water and sediment. The Heihe River
8 draining the Qilian Shan in the NE Tibetan Plateau acts as a dominant sediment routing
9 system from the uplifted NE Tibetan Plateau to the Hexi Corridor as well as the Badain
10 Jaran Desert. Reconstructing the evolution of the Heihe River could provide evidence for
11 the birth of the Badain Jaran Desert and enhance the understanding of sedimentary basin
12 fill and the relationship between tectonism, drainage evolution and environmental
13 changes. With this aim, two parallel cores (DWJ and XKJD with depths of 140 and 68.2
14 m, respectively) were drilled in the floodplain of the Heihe River. The facies analysis of
15 the sedimentary sequences from the drilling cores showed that the sedimentary
16 environment changed from the lake system to a delta system and finally to a fluvial
17 system at the depths of ~133.3 and ~68 m, respectively. The magnetostratigraphic results
18 revealed ages of approximately 1.75 and 1.12 Ma for the DWJ and XKJD cores,
19 respectively, and an age of approximately 1.1 Ma for the transition from delta to fluvial
20 environment in both the cores. The change of the sedimentary environment at
21 approximately 1.1 Ma was caused by the formation of the integrated Heihe River. The
22 integrated Heihe River may have developed via mechanisms such as river capture and
23 river diversion due to the uplifting of the North Qilian Shan and the Longshou Shan. The
24 present study suggested that the formation of large inland rivers, such as the Heihe River

and the Shiyang River in NW China, played an important role in the evolution of the deserts and the Loess Plateau.

Keywords: Badain Jaran Desert; Drainage evolution; Drill cores; Magnetostratigraphy; Heihe River

1. Introduction

The uplifting of the Tibetan Plateau, which was driven by the Indian–Eurasian collision, was one of the most predominant geological events during the Cenozoic Era (Molnar and Tapponnier, 1975; Li et al., 1979; Molnar and England, 1990; Tapponnier et al., 2001). The uplifting of the Tibetan Plateau has caused significant changes in environmental regimes in the surrounding regions (Harrison et al., 1992; Li et al., 1999; An et al., 2001; Molnar, 2005). The influences of the plateau on the environment were not only through its high elevation as a barrier to the atmospheric circulation (Manabe et al., 1990; Kutzbach et al., 1993) but also through the increasing denudation rates that produce abundant sediments which are delivered to the surrounding regions and the regions far from the plateau (Stevens et al., 2013; Bird et al., 2015; Wang et al., 2015; Zheng et al., 2015).

Drainage systems are the key tools in transporting sediment denuded in the Tibetan Plateau to the region where the desert forms and the loess material is emitted (Smalley et al., 2009; Yang et al., 2011; Stevens et al., 2013; Nie et al., 2015). Drainage evolution, which is supposed to be a mechanism of episodic sediment delivery, alternative to tectonic or climatic episodicity (Sobel, 2003), would greatly constrain the evolution of

47 sand seas and the Loess Plateau (Pell et al., 1997; Hollands et al., 2006; Vermeesch et al.,
48 2010; Yang et al., 2012; Li et al., 2014; Stevens et al., 2013; Bird et al., 2015; Wang et al.,
49 2015; Nie et al., 2015).

50 In NW China, the two large deserts, Badain Jaran and Tengger are located downstream
51 of the Heihe River and the Shiyang River, respectively. Both the Heihe River and the
52 Shiyang River originate from the NE Tibetan Plateau (Fig. 1). The sediment transported
53 by the Heihe River and the Shiyang River was a major source for the Badain Jaran Desert
54 and the Tengger Desert (Yang et al., 2012; Li et al., 2014; Wang et al., 2015; Hu and
55 Yang, 2016) and was also argued to be one of the dust sources for the Chinese Loess
56 Plateau (Derbyshire et al., 1998; Che et al., 2013). A previous study suggested that the
57 Shiyang River formed approximately 1.2 Ma (Pan et al., 2007), based on studies of the
58 highest fluvial terraces. This age is consistent with the formation age of the Tengger
59 Desert (Li et al., 2014). A recent study in the Badain Jaran Desert (Wang et al., 2015)
60 suggested a formation age of at least ~1.1 Ma based on electron spin resonance (ESR)
61 dating of aeolian sands from a 310 m drilling core, whereas scarce evidence has been
62 documented regarding the formation age of the Heihe River.

63 By regional strata comparison, Cao (1996) found that the modern Heihe River drainage
64 system was established between 1.05 and 0.36 Ma. Based on ESR dating on the highest
65 terrace of the upstream of the Heihe River, Tan et al. (1998) suggested that the Heihe
66 River formed at least 1.5 Ma ago but did not drain through the Jiudong Basin (Fig. 2)
67 until the Middle Pleistocene. Despite several focused studies, the evolution of the Heihe
68 River remains poorly constrained, which will greatly impede the understanding of the

69 environment evolution in NW China.

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71

72 **Fig. 1.** Map showing the distribution of river, loess and desert in the NE Tibetan Plateau, and patterns of
73 large-scale atmospheric circulation. The main map corresponds to the area within the rectangle in the inset map.
74 The desert distribution data are provided by the Environmental and Ecological Science Data Center of West China,
75 National Natural Science Foundation of China (<http://westdc.westgis.ac.cn>). The area of the Loess Plateau is
76 modified from Li et al., 2014.

77

78 To constrain the formation age of the integrated (integrating several rivers draining the
79 mountain range into a single river in the mountain front) large inland river, one of the best
80 possibilities is to find evidence from the mountain front (the Hexi Corridor, Fig. 2) where
81 tributary rivers that originated in the mountain region merge into one large river. The
82 sediment preserved in the Hexi Corridor (Fang et al., 2005) provides a record for
83 establishing the drainage evolution of the Heihe River. This study aimed to analyze the
84 sedimentary facies and paleomagnetic results of two drilling cores from the mid-stream of

the Heihe River in the Jiudong Basin (Fig. 2). Based on detailed sedimentary facies and magnetostratigraphic analysis, the present studies provide absolute age control and sedimentary framework related to the evolution of the Heihe River.

Fig. 2. Catchment of the Heihe River (shaded relief map based on 90 m SRTM data). (The boundary of the catchment is provided by Cold and Arid Regions Sciences Data Center at Lanzhou, <http://westdc.westgis.ac.cn/>). Triangle a: Yumu Shan; Triangle b: Wenshu Shan.

2. Geographic and geological setting

95 The Heihe catchment, which ranges between 37°45' and 43°10' N and between 96°8'
96 and 104°8' E, is located at the intersection of the NE margin of the Tibetan Plateau and
97 the southern slopes of Gobi-Altay Shan (Figs. 1 and 2). The catchment hosts the second
98 largest continental alluvial fans in the world and the third largest desert (the Badain Jaran
99 Desert) in China (Fig. 2). The Heihe catchment covers an area of 271,000 km² according
100 to the boundaries delineated in 2010, by West Data Center (West D C) of the Cold and
101 Arid Regions Environmental and Engineering Research Institute, Chinese Academy of
102 Sciences (West D C, 2011a) (Fig. 2). The Heihe catchment has 41 perennial tributaries
103 (West D C, 2011b) and they all originate from the Qilian Shan. The Heihe River is the
104 second largest endorheic river in China with a total length of 812 km and a runoff volume
105 of $15.8 \times 10^8 \text{ m}^3$ per year, which accounts for about half of the total runoff of all the rivers
106 in the catchment. The Heihe River drains the Qilian Shan, flows through the Hexi
107 Corridor and finally disappears into the Juyan Lake (Fig. 2). The Heihe River can be
108 divided into three segments from south to north by the Yingluo Gorge and Zhengyi Gorge
109 (Fig. 2). The segment up to the Yingluo Gorge is treated as the upstream area, located in
110 the mountain area of the Qilian Shan, with an elevation of ~4000 m. The annual
111 precipitation in the upstream area of the Heihe River is over 300 mm (Yao et al., 2014).
112 Between the Yingluo Gorge and the Zhengyi Gorge, the midstream area of the Heihe
113 River is located in the Hexi Corridor, with elevations between 1500 and 1700 m (Fig. 2).
114 The annual precipitation in the midstream areas of the Heihe River is approximately
115 80–130 mm (Yao et al., 2014). Below the Zhengyi Gorge, the downstream area of the
116 Heihe River is located in the stable Alashan block, with elevations between 900 and 1200

m. The mean annual precipitation in the downstream area of the Heihe River is ~40 mm (Yao et al., 2014).

The Hexi Corridor (Fig. 2) is an NWW elongated Cenozoic foreland basin developed on a Jurassic foreland basin (Wang and Coward, 1993) or on an Early Cretaceous extensive basin (Gansu Geologic Bureau, 1989). It is bounded by the Qilian Shan to the south and the Jinta Nan Shan–Heli Shan–Longshou Shan to the north (Gansu Geologic Bureau, 1989; Fig. 2). The Heihe River flows along the northern margin of the Jiudong Basin (Fig. 2). The Jiudong Basin is located in the western part of the Hexi Corridor, and constrained by the Yumu Shan to the east and the Wenshu Shan to the west (Liu et al., 2011; Fig. 2). Biostratigraphic and magnetostratigraphic results indicate that the Paleocene and Eocene strata are absent in the Jiudong Basin and that the Cenozoic deposition began in the Oligocene (Fang et al., 2013). The maximum thickness of the Quaternary deposits is over 1000 m in the depocenters of the Jiudong Basin (Editorial Group of Petroleum Geology of Yumen Oilfield, 1989). Quaternary deposits in the Jiudong Basin comprise dominantly of glaciofluvial, alluvial, fluvial–lacustrine, and aeolian facies (Cao, 2010; Fig. 3). The Pleistocene sequence in the southern margin of the Jiudong Basin is composed of the Yumen Formation (Yumen conglomerates) and the Jiuquan Formation (Jiuquan gravels) (Fang et al., 2005). The Yumen Formation (Q_1) consists of grayish brown beds of glaciofluvial, diluvial conglomerate and lenticular sandstone (Fang et al., 2005; Zhu et al., 2006; Liu et al., 2010). The Jiuquan Formation (Jiuquan gravels, Q_2 – Q_3), which unconformably overlies the Yumen conglomerates, consists of gray black beds of glaciofluvial, diluvial gravel, sandy gravel, banded sand

139 and clay (Zhao et al., 2001; Fang et al., 2005; Zhu et al., 2006). The Holocene sequence
140 mainly consists of diluvial, alluvial, fluvial–lacustrine, and aeolian sediments (Gansu
141 Geologic Bureau 1989; Fig. 3).

142 Previous research (Wang, 1990) showed that the lower part of the Quaternary sequence
143 in the central–northern Jiudong Basin is predominantly composed of fluvial–lacustrine
144 deposits, which are named the Bageleng Formation. The Bageleng Formation is
145 considered to be a heteropic product synchronous with the Yumen conglomerates (Wang,
146 1990; Zhu et al., 2006). The Bageleng Formation is composed of partially cemented
147 mudstone, siltstone, and pebbly sandstone (Wang, 1990; Zhu, 2006).

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Fig. 3. Simplified geologic map of the research area and the locations of the drilling cores (DWJ and XKJD). The map is modified from the 1:500,000 geologic map of China. NQF – Northern Qilian fault, NYF – northern Yumu Shan fault, SHF – southern Heli Shan fault. The North Qilian Shan and the Yumu Shan dominantly consist of pre-Mesozoic rocks (Gansu Geologic Bureau, 1989). The Heli Shan is composed mainly of Proterozoic rocks, Mesozoic rocks and Variscan granite (Gansu Geologic Bureau, 1989).

3. Materials and Methods

3.1 Drilling and Sampling

Two parallel cores (named DWJ and XKJD, at a horizontal distance of 1.5 km) were drilled on the southwest fluvial plain of the Heihe River in the Hexi Corridor (Fig. 3)

using a XY-2200 m machine drilling system with a single pipe sampling system. The drilling depth of DWJ (39°29'26.4" N, 99°36'19.2" E, Elevation: 1338 m) is 140 m, and the length of the recovered core is 125.25 m, with a recovery rate of 89.5%. The drilling depth of XKJD (39°29'51.2" N, 99°37'21.5" E, Elevation: 1341 m) is 68.2 m, and the length of the recovered core is 57.65 m, with a recovery rate of 86.2%.

The recovered sediments consist mostly of clay, silt, and fine sand, whereas medium to coarse sand and gravels cannot be collected completely during drilling because they are unconsolidated and easy to fall out from the barrel. Unconsolidated sediments were described in the field, while compacted and cemented core sections were described in the laboratory. Sediments description including sediment colors, sedimentary textures and sedimentary structures was done by visual observation according to Miall (2000). Color was described using a fresh surface and comparisons to the National Research Council Rock-Color Chart (Goddard et al., 1948). Lithologic classification was based on grain size. For sand-sized sediment, a grain size chart was used for distinguishing the main sand-sized sub-divisions: very coarse, coarse, medium, fine, very fine. Such observations could provide adequate information on the modal size range of the sand layers. The description was modified by appropriate adjectives if sorting characteristics require it; for example, pebbly coarse sand, silty clay, etc. Silt and clay were distinguished in a hand specimen by the presence or absence of a gritty texture, as felt by the fingers or the tongue. For gravels, maximum clast size was typically estimated by taking the average of the ten largest clasts visible in the drilling core. It is also important to note the degree of sorting, clast shape and roundness, matrix content. Oriented samples for magnetism

measurement were taken at 0.1–1 m intervals (except for a few samples at intervals of 2–6 m) and then cut into 10 cm thick slices perpendicularly to the core axis. A total of 216 oriented samples from the DWJ core and 100 oriented samples from the XKJD were taken for paleomagnetic measurements. In the field or the lab, two or three cubic specimens of 2×2×2 cm were subsequently cut from each sample.

3.2 Paleomagnetic analysis

3.2.1 Rock magnetism measurement

Paleomagnetic chronology is based on separating of the ChRM (characteristic remanent magnetization). Separating of the ChRM carried by different magnetic minerals relies on different demagnetization methods and steps. The study of rock magnetism provides bases for identifying the categories of magnetic minerals in the sediments (Dunlop and Özdemir, 1997). In order to choose the appropriate demagnetization methods and steps, we carried out the rock magnetism measurement. Fifteen specimens of different lithology from the DWJ and XKJD cores with intervals of 10–20 m were selected for the measurement of the isothermal remnant magnetization and thermomagnetic analysis. Isothermal remnant magnetization was measured in a Molspin Minispin rotation magnetometer using a maximum external magnetic field of 1 T. The temperature dependence of the magnetization was measured using a variable field translation balance (VFTB) from room temperature (~20 °C) to 700 °C and then back to room temperature, with an interval of 4 °C in a 110 mT applied field.

3.2.2 Demagnetization of the Natural Remanent Magnetization

Thermal demagnetization will be better than AF demagnetization for some of our samples because contributions from hematite cannot be removed by AF demagnetization. However, specimens in the upper part of the drilling cores are too loose for thermal demagnetization. AF demagnetization was employed for all the samples and thermal demagnetization was employed for the parallel samples with good cementation. Remanent magnetization was measured using a 2G Enterprise 760 Cryogenic Magnetometer installed in field-free space (<200 nT). 316 oriented specimens were subjected to stepwise alternating-field (AF) demagnetization at fields up to 80 mT with 5 or 10 mT increments. 138 specimens were subjected to progressive thermal demagnetization up to a maximum temperature of 585°C or 685°C with 25–50°C intervals below 610°C and 10–15°C above 610°C, using a MMTD80 thermal demagnetizer.

Demagnetization results were evaluated using orthogonal projection (Zijderveld, 1967) diagrams. ChRM (Characteristic remanent magnetization) directions were calculated using principal component analysis (PCA) of the least-squares method (Kirschvink, 1980). All of the paleomagnetic data were analyzed using Paleomagnetism data analysis version 4.2 (PMGSC 4.2) developed by R. Enkin (http://gsc.nrcan.gc.ca/sw/paleo_e.php).

Rock magnetic and paleomagnetic experiments were conducted at the Paleomagnetic and Rock Magnetic Laboratory of the Key Laboratory of Western China's Environmental Systems (Ministry of Education of China) at Lanzhou University.

4. Results

4.1. Lithology

The sedimentary sequence of the two cores comprised the following lithological units, from bottom to top (Fig. 4).

Unit 1: 140–133.2 m (DWJ, Fig. 4). This Unit predominantly consisted of yellowish, grayish yellow silty clay. The silty clay layer was intercalated with brownish yellow thin-bedded clayey silt. This Unit contained some carbonate concretions, phytoclasts, and horizontal lamination.

Unit 2: 133.6–69.7 m (DWJ) and 68.2–65.6 m (XKJD) (Fig. 4). This unit mainly consisted of silt and sand, and could be divided into three lithological subunits:

(i) 133.2–106.3 m (DWJ, Fig. 4). In this subunit, brownish yellow thin- to medium-bedded silt were intercalated with grayish yellow thin-bedded clay and yellow thin- to medium-bedded fine-grained sand. The silt layers were generally massive or weakly laminated. The clay layers are massive, laminated and occasionally wavy-bedded. Both silt and clay layers contained numerous carbonate concretions.

(ii) 106.3–84.8 m (DWJ). This subunit consisted mainly of grayish yellow, grayish green and reddish medium- to thin-bedded medium and fine-grained sand. The sand layers were intercalated with reddish, yellowish thin-bedded silt and clay. Sand layers were generally 1–3 m thick. Some of the sand layers were partially cemented by calcium carbonate. Sand layers in 98.9–98.5 m and 110.1–109.9 m contained gypsum. In some horizons, sand layers (96.5–95.4 m and 90.7–90.5 m) were red and contained iron–manganese plaque. The bottom of the sand layers contained reddish mud lumps and

medium-sand-sized mud balls. Very fine sand layers were usually parallel-laminated. The silt and clay layers were generally massive and occasionally laminated. Some of those layers contained carbonate concretions, black carbonized rootlets, phytoclasts, and occasionally dwelling burrows.

(iii) 84.8–69.7 m (DWJ), 68–65.6 m (XKJD) (Fig. 4). This subunit consisted of alternating grayish green, grayish yellow and reddish thick-bedded silt, clay and grayish green, grayish yellow thin- to medium-bedded fine-grained sand. The silt and clay layers were generally massive or laminated, and several layers contained carbonate concretions, black carbonized rootlets, phytoclasts, and a few dwelling burrows. The sand layers occasionally showed parallel bedding, and some sand layers were partially cemented by carbonates.

In the XKJD core, the interval of 68.2–65.6 m consisted of brownish yellow, grayish yellow silt and silty clay. These layers contained some carbonate concentrations. This sequence was correlated to the upper part of the unit 2 in the DWJ core.

Unit 3: 69.7–0 m (DWJ), 65.6–0 m (XKJD) (Fig. 4). This unit contained several thick layers of grayish yellow, yellowish sandy gravel, pebbly sand. These coarse-grained layers were intercalated with yellowish, grayish green, thin- to medium-bedded very fine sand, clayey silt, and silty clay. The gravel layers in this unit were mainly clast-supported. Some gravel layers showed matrix support and contained intraclasts at the bottom. Concave erosion surfaces were present in the boundary of some deposition cycles. Gravels were subrounded–rounded. The maximum clast size of the gravels was 3 cm and 4 cm in the DWJ and XKJD cores, respectively.

Fig. 4. Lithology, sedimentary structure and interpreted sedimentary facies of the DWJ and XKJD drilling cores. c

— clay, sl — silt, vfs — very fine sand, ms — medium sand, vcs — very coarse sand, p — pebble.

4.2. Rock magnetic and paleomagnetic results

4.2.1. Rock magnetic results

The isothermal remnant magnetization (IRM) acquisition curves and demagnetization curves are shown in Fig. 5. The IRM of pilot samples reaches 80%–93% of their SIRM

under 300 mT (Fig. 5), indicating that most of the remnant magnetism is carried by ferrimagnetic minerals (Liu et al., 2003). The IRM for some specimens continue to increase after 300 mT (Fig. 5), indicating that those specimens contain high-coercivity magnetic minerals (Nowaczyk, 1997).

Fig. 5. Normalized isothermal remanent magnetization (IRM) acquisition curves (right) and their reverse field demagnetization curves (left) for the pilot samples. The IRM curves are normalized to the saturated IRM acquired at 1000 mT, defined as SIRM.

The sharp decrease in the magnetic susceptibility at $\sim 580^{\circ}\text{C}$ in the thermomagnetic curves (Fig. 6) reveals the existence of magnetite. The sustained decrease in the magnetic susceptibility with several samples between 600°C and 680°C (Fig. 6) indicates the existence of hematite in these samples. The slight decrease in the magnetic susceptibility between 250°C and 500°C (Fig. 6) indicates the existence of small amount of maghemite

(Dunlop and Özdemir, 1997). No changes of magnetic susceptibility between 0°C and 120°C are observed (Fig. 6), indicating the absence of the goethite (Dunlop and Özdemir, 1997). In summary, the synthetic rock magnetic experiments suggest that sediments contain magnetite, hematite and small amount of maghemite.

Fig. 6. Thermomagnetic curves (magnetic susceptibility versus temperature) for samples from the DWJ and XKJD cores.

4.3.2. Paleomagnetic Results

Representative demagnetization diagrams are shown in Fig. 7. AF and Thermal demagnetization isolated several distinct directions. A low field or low temperature component (softer magnetite particles or a viscous overprint) was removed below ~30mT or ~300°C. The characteristic remanent magnetization (ChRM) component was separated between 30 mT and 80 mT (e.g., Fig. 7a, 7b, 7c, 7d, and 7e) or between 300–350°C and 585°C (e.g., Fig. 7f) for some specimens. However, for some specimens, which could not be fully demagnetized by AF demagnetization, the high-stability ChRM component persists up to 685°C (e.g., Fig. 7h, 7j, and 7l). As the magnetic directions are similar at 550–585°C and 650–685°C in those samples (e.g., Fig. 7h, 7j, and 7k), the lock-in time of the remanence in both magnetite and hematite likely occurred during or very soon after

deposition. The ChRM directions for these specimens which were determined using the points above 610°C are similar to that determined using the points between 400°C and 585°C, and similar to that determined by AF demagnetization (e.g., Fig. 7g, 7h, 7i, 7j, 7k, and 7l).

Fig. 7. Orthogonal demagnetization diagrams of the typical samples from the drilling cores (XKJD and DWJ). The closed (open) symbols represent the horizontal (vertical) projections. AF demagnetization levels are marked with mT (a, b, c, d, e, g) and the thermal treatment levels are marked in °C (f, h, j, l). NRM is the natural remanent magnetization.

The ChRM directions were calculated using principal component analysis (PCA) of the least squares method (Kirschvink, 1980). Reliable ChRM directions are established based on: (i) ChRM directions that could be determined without ambiguous or noisy orthogonal demagnetization diagrams, (ii) at least three successive demagnetization steps, (iii) maximum angular deviation (MAD) less than 15°. Among the 454 specimens, a total of

specimens (262 specimens were demagnetized by AF method and 79 were demagnetized by thermal method) gave reliable ChRM directions. From the analyzed ChRM directions, only the inclination is valuable because there was no control on the core azimuth during drilling. Therefore, the magnetic polarity sequences were constructed based only on the inclination values of the DWJ and XKJD cores.

The mean positive and negative inclinations were calculated by the Arason–Levi method based on the maximum likelihood estimation (Arason and Levi, 2010). AF demagnetized specimens in the DWJ core (Table 1) yield a mean normal inclination of $I = 52.14^\circ$ and a mean reversed inclination of $I = -49.1^\circ$. AF demagnetized specimens in the XKJD core (Table 1) yield a mean normal direction of $I = 44.83^\circ$, and a mean reversed ChRM direction of $I = -41.85^\circ$. Thermally demagnetized specimens have a slightly different direction population (may be result of less quantity): The directions isolated by thermal demagnetization in the DWJ core have a mean normal direction of $I = 52.34^\circ$ and reversed direction of $I = -54.61^\circ$. The directions isolated by thermal demagnetization in the XKJD core have a mean normal direction of $I = 46.09^\circ$ and a mean reversed direction of $I = -48.02^\circ$. The accepted inclinations (AF demagnetization) are shown in Fig. 8 with an interval of 3° . The distribution patterns of inclinations show that both the positive and the negative inclinations follow normal distributions. In both cores, the difference between the normal and the reversed inclinations is less than 5° , suggesting that they are antipodal and pass the reversal test (McFadden and McElhinny, 1990). Therefore, the measured inclinations for the DWJ and XKJD cores are reliable for reconstructing the polarity sequences.

Table 1

Statistic paleomagnetic results from DWJ and XKJD drilling core.

			N	I (°)	k	α_{95} (°)
AF Thermal demagnetization	DWJ	Postive-incl.	52	52.14	14.78	5.32
		Negative-incl.	133	49.1	13	3.53
	XKJD	Postive-incl.	44	44.83	12.47	6.35
		Negative-incl.	33	41.85	11.74	7.62
	DWJ	Postive-incl.	14	52.34	9.15	8.75
		Negative-incl.	43	54.61	8.73	4.56
	XKJD	Postive-incl.	8	46.09	8.95	12.63
		Negative-incl.	14	48.02	10.39	10.36

N – number of samples, I – mean inclination, k – precision parameter, α_{95} – 95% confidence limits.

Fig. 8. Histogram of the reliable inclinations of the drilling cores (DWJ and XKJD) with 3° windows (AF demagnetized samples).

Following rules that the single dating point was ignored in the final interpretation, at least two successive dating points with a same polarity were considered to be an excursion or event (Li et al., 2014), we determined four normal and four reversed polarity zones (marked DN1–N4, DR1–R4) in the DWJ core and determined two normal and two

reversed polarity zones (marked XN1–N2, XR1–R2) in the XKJD core (Fig. 9).
Determined polarity results by AF demagnetization are in good agreement with the results
by thermal demagnetization (Fig. 9).

5. Data interpretation and discussions

5.1. Analysis of sedimentary facies (environments)

The sedimentary sequence of the drilling cores exhibited an overall coarsening-upward trend. Based on the lithology, sedimentary texture, structure and their vertical associations, as well as sedimentary models (Reading, 1996), the sequence could be divided into three different sets of facies that represent three different sedimentary environments.

Lake: 140–133.2 m (Unit 1) (Fig. 4). This unit consisted of silty clay and contained horizontal lamination. Silty clay was mainly grayish yellow. This sequence was formed under calm conditions and might be of shallow-water origin (Reading, 1996). This unit was interpreted as lacustrine facies. The shallow lake facies are commonly associated by the floodplain and deltaic sequence (Reading, 1996).

Delta: 133.6–69.7 m (DWJ), 68.2–65.6 m (XKJD) (Unit 2) (Fig. 4). This unit consisted of silt and sand, and exhibited an overall coarsening-upward characteristic. This unit was interpreted as delta facies. It contained three subfacies.

(i) Delta front: 133.2–106.3 m (Fig. 4). This sequence had a gradational contact with deposits of the shallow lake. It contained two medium-scale (10–20 m) progradational sequences (133.6–120 m and 120–106.3 m), which showed an upwards gradation from clay, silt to sand. The transition from silt to sand layer indicated a marked shallowing of

the depositional environment. The fine-grained clay, silt were deposited during periods of high lake level. The sand layers were deposited during periods of low lake level, when delta facies prograded into the lake (Dam and Surlyk, 1993). Higher in the sequence, thin sand layers were intercalated with parallel silt laminae, reflecting a combination of sediment-laden current incursion from the distributaries and continued deposition from suspension (Reading, 1996). This sequence might be produced by mouth bar/distal bar progradation into the lakes (Coleman and Wright, 1975; Reading 1996). This sequence was interpreted as delta front subfacies.

(ii) Delta plain: 106.3–84.8 m (Fig. 4). This sequence contained four fining-upward sedimentary cycles. This sequence was capped by clay beds with associated rootlets and wave-lamination deposits (Fig. 4; sub-unit iii in unit 2) indicative of delta environment (Reading, 1996). It began with medium sand of lag deposit and reddish mud balls of intraclasts and was interpreted as the product of distributary channel sedimentation (Dam and Surlyk, 1993). These sand/gravel-size mud aggregates might have been eroded by floods and transported as bedload after the underlying fine-grained sediments were deeply cracked (Rust and Nanson, 1989). In some horizons, sand layers (96.5–95.4 m and 90.7–90.5 m) were red and contained iron–manganese plaques, indicating that the sand layers were exposed and weathered after they were deposited (Reading, 1996). Sets of the sand layers, containing intraclasts and exposure markers, might be produced by ephemeral distributary channel swinging in the delta plain (Cant and Walker, 1978; Reading, 1996; Davies and Elliott, 1996). This sequence was interpreted as delta plain subfacies.

(iii) **Delta front:** 84.8–69.7 m (Fig. 4). This subunit contained two progradational sequences (84.8–80.7 m and 80.7–69.7 m), which coarsened upward from clay, silt to sand. This sequence began with clay beds with associated rootlets and wave-lamination deposits (Fig. 4) indicative of delta environment. Higher in the sequence, sand layers were intercalated with parallel and wavy silt laminae, reflecting a combination of sediment-laden current incursion from the distributaries and continued deposition from suspension (Reading, 1996). This sequence might be produced by mouth bar/distal bar progradation in the lakes during river floods (Reading, 1996). This subunit was interpreted as delta front subfacies.

River: 69.7–0 m (DWJ), 65.6–0 m (XKJD) (Unit 3) (Fig. 4). This unit contained three fining-upward sedimentary cycles, which graded upward from sandy gravels, pebbly sand to very fine sand, silt and clay. The thickness of each fining-upward cycle was approximately 15–20 m. The gravel layers in this unit were mainly clast-supported. Gravels were subrounded–rounded. These gravel layers recorded bedload deposition from stream flows (Nemec and Postma, 1993). This sequence was interpreted as fluvial facies (Miall, 1996; Reading, 1996).

In summary, sedimentary environment in the study area changed from lake (140–133.2 m in DWJ core) to delta (133.6–69.7 m in DWJ core, 68.2–65.6 m in XKJD core) and finally shifted to river (69.7–0 m in DWJ core, 65.6–0 m in XKJD core).

5.2. Age constraints on the sedimentary sequence

The reconstructed polarities sequences (Fig. 9) from DWJ and XKJD cores can be correlated with the standard geomagnetic polarity time scale (GPTS) (Gradstein et al., 2004; Laj and Channell, 2007) (Fig. 9). The topmost predominantly normal polarity zones, DN1 (0–47 m) in DWJ core and XN1 (0–40 m) in XKJD core (Fig. 9), can be correlated to Chron C1n (the Brunhes Normal Polarity Epoch). Magnetozone XN2 (59.1–61 m) in the XKJD core is correlated with Chron C1r.1n. The magnetozone DN2 of DWJ core may correlate to Chron C1r.1n or the Kamikatsura excursion event. The short normal polarity zone DN3 (96.8–98.5 m) and DN4 (131.7–133.1 m) within the Matuyama Reverse Polarity Epoch may correlate to Chron C1r.2n and C1r.3n or C1r.1n and C1r.2n. Thus, we propose two possible correlations of the recognized magnetozones to the GPTS (Fig. 9). We prefer correlation I because correlation II requires more unrealistic variations of sediment accumulation rates (SAR) (Fig. 10): (i) we note that unit 3 in the DWJ core and the XKJD core have a similar thickness, both of those sequence belong to fluvial facies (Fig. 9). Those two sequence deposited in the same tectonic setting may have a similar sediment accumulation rate. But correlation II gives an anomalously high SAR relative to that in the core XKJD as well as the upper part of unit 3 in the DWJ core (Fig. 10); (ii) Correlation II gives extremely low values in magnetozone DN3 relative to DR2 and DR3 (Fig. 10). However, those sections show no large changes in sedimentary facies.

Fig. 9. Magnetostratigraphic results of the DWJ and XKJD cores and their correlation with the geomagnetic polarity timescale (GPTS) (Gradstein et al., 2004; Laj and Channell, 2007). Red circles and lines are thermal demagnetization results, and black circles and lines are AF demagnetization results. Small round symbols (not connected by lines with neighboring data) represent single samples within otherwise opposite polarity zones. Correlation I is preferred in this study.

Consequently, the age of the DWJ core and the XKJD core was constrained to ~1.75 and ~1.12 Ma, respectively. According to the sediment accumulation rate, the deltaic sedimentary environment in the DWJ core lasted from 1.68 Ma to 1.1 Ma. The bottom age of the fluvial facies in the DWJ and XKJD cores was estimated to be approximately 1.1 and 1.08 Ma, respectively.

Fig. 10. Age versus depth plot showing calculated linear average sediment accumulation rate (SAR) using correlations I and II of Fig. 9.

5.3. Reconstructing drainage evolution in the Jiudong Basin

5.3.1. Drainage systems before 1.1 Ma

The analysis of the sedimentary facies and magnetostratigraphy showed that lakes existed in this area before 1.68 Ma (Fig. 11a). The lakes shrank, and the delta system prograded after 1.68 Ma (Fig. 11b). The shift of the sedimentary environment was coeval with the formation of the modern drainage system of the Yellow River (Li et al., 1996). Based on sections 5.1 and 5.2, it was suggested that an ephemeral river existed in this area before 1.1 Ma. The ephemeral river may flow perpendicularly to the trend of the North Qilian Shan (Fig. 11b). The lacustrine clay of the Early Pleistocene with a

thickness of 200–300 m was found in the Nanhua Village (Fig. 3), east of the drilling cores (Institute of Geology and Lanzhou Institute of Seismology, 1993). This thick lacustrine clay indicated that the west-flowing longitudinal river (parallel to the trend of the North Qilian Shan) did not exist in the study area during the Early Pleistocene. In fact, fine-grained, fluvial–lacustrine facies was widespread across the Kuantan Shan and the midstream of the Heihe River during the Early Pleistocene (Wang, 1990; Institute of Geology and Lanzhou Institute of Seismology, 1993). The transverse river, which drained the Qilian Shan and flowed into ancient lakes in the Hexi Corridor, was similar to the present Mayinghe River (an ephemeral river, Fig. 3), which drained into the Yanchi lake (Fig. 3).

According to facies association relationship (Walther facies law), coeval alluvial fan sedimentation further south and lake sedimentation further north to the research area should exist during the deposition of unit 1 and unit 2. Early Pleistocene alluvial fan sedimentation (the Yumen conglomerates) is widespread in the piedmont of the Qilian Shan (Fang et al., 2005). Lakes were reported in the northern Hexi Corridor during the Early Pleistocene (Wang, 1990; Institute of Geology and Lanzhou Institute of Seismology, 1993). Sedimentary facies changed from alluvial fan facies in the piedmont of the Qilian Shan to the fine-grained, fluvial–lacustrine facies (the Bageleng Formation) in the central–north of the Jiudong Basin during the Early Pleistocene (Editorial Group of Petroleum Geology of Yumen Oilfield, 1989; Wang, 1990; Cao, 2010; Liu et al., 2010). The variation trend of the sedimentary facies across the Jiudong Basin together with the drilling data indicate that the drainage system before 1.1 Ma in the Jiudong Basin might

497 consist of alluvial fans in the piedmont, transverse rivers, deltas, and lakes in the
498 central–north of the Jiudong Basin, as shown in Fig. 11a and Fig. 11b.

(a) >1.7 Ma

(b) $1.7-1.1$ Ma

(c) <1.1 Ma

499

500 **Fig. 11.** Drainage patterns/basin filling models in the Jiudong Basin (a) before 1.7 Ma, (b) 1.7 – 1.1 Ma, and (c)

501 1.1 Ma till present.

502

503 *5.3.2. Emergence of the modern Heihe River*

504 A major shift of the sedimentary environment in the study area from the lake–delta
505 system to a fluvial system occurred at approximately 1.1 Ma. However, it is not a
506 common phenomenon in the Jiudong Basin as lakes persist in the Yanchi (Li et al., 2013)
507 (Fig. 3). According to the magnetostratigraphic chronology (Fig. 9) and the sediment
508 accumulation rate (Fig. 10), the upper succession (84.8–69.7 m; subunit iii; Fig. 4) of the
509 delta system formed approximately between 1.2 Ma and 1.1 Ma. This succession
510 composed of two coarsening-upward sedimentary cycles (section 5.1; Fig. 4), and the
511 sedimentary periodicity in this succession could be estimated to as ~50 ka. Based on the
512 magnetostratigraphic chronology (Fig. 9), the bottom age of the three sedimentary cycles
513 in unit 3 (Fig. 4) for the DWJ and XKJD cores was placed at about 1.1, 0.82, 0.59 and
514 1.08, 0.81, 0.51 Ma, respectively, and thus the sedimentary periodicity in unit 3 could be
515 estimated to as ~300 ka. Sedimentary periodicity recorded in the present sedimentary
516 sequence changed from ~50 ka to ~300 ka at ~1.1Ma. This time was prior to the time
517 when the major shift in the dominant climatic periodicity took place (Lisiecki and Raymo,
518 2005). The major shift recorded in the marine $\delta^{18}\text{O}$ from 41 ka to 100 ka was at about
519 1–0.8 Ma (Ding et al., 2002; Lisiecki and Raymo, 2005). Climatic proxy records from the
520 studied basins in the NE Tibetan Plateau indicate an enhancement of aridification during
521 1.2–0.6 Ma (Li et al., 2014). Against the background of enhanced aridification and
522 assuming that fluvial sediments in unit 2 and 3 were deposited by the same river, grain
523 size in unit 3 would be smaller than that in unit 2 because of a reduction in water

discharge. However, our data seem to indicate the contrary (Fig. 4). Changes in the sedimentary cycles and grain size in the DWJ and XKJD cores (Fig. 4) were not unique to the climate records of the Pleistocene but might be controlled by other factors, such as tectonism or drainage reorganization (Reading, 1996; Sobel et al., 2003; Clift and Blusztajn, 2005).

Previous research suggested that during times of tectonism and uplift in the thrust belt, coarse alluvial deposits accumulated only in the most proximal part of the basin and graded rapidly into fine-grained deposits toward the distal part of the basin (Heller et al., 1988). Coarse sediment in unit 3, which deposited in the distal part of the Hexi Corridor, could not be the immediate result of the tectonism and uplift in the North Qilian Shan (Li et al., 2014). The onset of the Heihe River in the north distal of the Jiudong Basin, which changed the configuration of the drainage system at ~1.1 Ma (Fig. 11c), may account for the shift from the lake-delta system to the stable fluvial system accompanied by a significantly coarser grain size.

The sedimentary characteristics of unit 3 showed significant differences compared with those of Unit 2: (i) the grain size of unit 3 was significantly coarser than that of unit 2; (ii) a thick layer of gravel (Fig. 4) appeared for the first time in the drilling core; (iii) sandy gravel layers appeared frequently in unit 3; (iv) the thickness of the sedimentary cycles in unit 3 is significantly increased compared with that in unit 2. All of the apparent changes in the sedimentary characteristics in unit 3 indicated that the stream power was significantly enhanced. The thickness of the sedimentary cycles in unit 3 was the same as that of the modern Heihe River in the Jiudong Basin (unpublished data). The gravels in

unit 3 were subrounded–rounded, with complicated lithology types, which were similar to the modern sediments of the Heihe River, indicating a connection between the unit 3 sediments and the Heihe River. Furthermore, except for the large Heihe River, no other rivers in this area had the ability to transport gravels from the Qilian Shan to the study area. According to the hydrogeological profile along the Mayinghe River, the Pleistocene sedimentary strata along the Mayinghe River can be divided into three successive belts from south to north (Zhang, 2005): the southern single layered belt, which consists of alluvial–diluvial sand and gravel; the central double layered belt, which consists of interbedded clay, silt, sand and gravel; and the northern multilayered belt, which consists of fine-grained clay, silt and sand. The Pleistocene sedimentary strata along the Bailanghe River had a similar distribution pattern with that of the Mayinghe River (Institute of Geology and Lanzhou Institute of Seismology, 1993; Zhang, 2005). The drilling cores are located in the north of the fine-grained multilayered sedimentary belt. Assuming that sediment in the study area was dispersed by a north-flowing river such as the Mayinghe River and Bailanghe River, it can be expected that sediment of unit 3 in the drilling cores might consist dominantly of clay, silt, and fine-grained sand. However, our data indicate that unit 3 consisted dominantly of coarse-grained gravel and sand (Fig.4), which is in significant contrary to the distribution of the sedimentary strata formed by the Mayinghe or the Bailanghe River. Sediment of unit 3 could not be dispersed by the north-flowing river. Based on the analysis above, it is suggested that unit 3 in the drilling cores was formed by the Heihe River.

The drainage system changed significantly after ca 1.1 Ma. Ancient lakes and deltas disappeared in the study area. The Heihe River, which had much greater carrying and scouring capacity, cut through the ancient lakes and flowed into the Ejina Basin. Previous independent transverse rivers in the study area turned into tributaries of the Heihe River (Fig. 11c).

5.4. Tectonic controls on drainage evolution of the Heihe River

Tectonic movement, which steepened the surface gradient (Clark et al., 2004) and might induce climatic changes (Liu et al., 2015), caused strong incision (Pan et al., 2007), headwater erosion, river capture, and river diversion (Brookfield, 1998) and finally induced reorganization of the previous drainage systems (Babault et al., 2012).

Previous research indicated the existence of a river flowing through the Yingluo Gorge during the Early Pleistocene (Li et al., 1995). The river might be the paleo-Heihe River, but had not yet flowed into the Jiudong Basin. Drilling data in Nanhua Village and Zhangye City (Fig. 3) showed that Early Pleistocene lakes existed in these areas (Institute of Geology and Lanzhou Institute of Seismology, 1993). It can be inferred that the paleo-Heihe River drained into a paleolake in the northern Hexi Corridor, which was similar to the drainage system in the Jiudong Basin during the same epoch. The drainage system changed significantly by achieving an enlarged catchment area at approximately 1.1 Ma, under the background of enhanced aridification (Li et al., 2014). The enlarged catchment area of the paleo-Heihe River might be a result of headwater erosion and river capture at the interval of tectonic uplift and compressional deformation in the North

Qilian Shan (Li et al., 2014). At the same time, the uplifting of the Longshou Shan (Institute of Geology and Lanzhou Institute of Seismology, 1993) forced the reorganized Heihe River to flow into the Jiudong Basin. Due to the northward propagation of the transverse fans in the Jiudong Basin resulting from asymmetric tectonic uplift in the North Qilian Shan and the Longshou Shan–Heli Shan (Li et al., 1998), the enlarged Heihe River was constrained at the north edge of the Jiudong Basin.

5.5. Implications for long-term landscape evolution in NW China

The Heihe River might have flowed into the Ejina Basin at approximately 1.1 Ma. Drilling data showed that the Ejina Basin was filled with fluvial–lacustrine sediments with a thickness of approximately 340 m (Wen et al., 2005; Yao et al., 2014). The deposition of the fluvial sediment in the Ejina Basin was roughly supposed to begin in the Early Pleistocene (Wang, 1990; Wen et al., 2005; Zhang, 2005). Sediments changed from gravel to sand, silt and clay across the basin from south to north (Wen et al., 2005), indicating that sediment deposited in these areas were south-derived. The river that could deliver gravels into the Ejina Basin had greater stream power. The river might be the Heihe River. The rivers originating from the Mazong Shan and the Jinta Nan Shan–Heli Shan did not have the ability to deliver gravels to the Ejina Basin due to the small catchment area and little precipitation (Yao et al., 2014). Based on the aforementioned, it was suggested that the Heihe River flowed into the Ejina basin at ~1.1 Ma.

The onset of the Heihe River in the Ejina Basin might have constrained the formation of the Badain Jaran Desert. A marked decrease in the sediment accumulation rate after 1.1

Ma (Fig. 10) might indicate that a large number of sediments were transported to the Ejina Basin by the Heihe River. Large volume of sediments built the second largest alluvial fan in the world. This large fluvial fan located upwind of the Badain Jaran Desert has been proved to be genetically related to the Badain Jaran Desert (Zhu et al., 1980; Yang et al., 2012; Hu and Yang, 2016). Under the dry climate regime and strengthened winter monsoon circulation after ~1.1 Ma (An et al., 2001), the clastic sediments in the fluvial fan were sorted by wind and were finally transported to the desert area. As a consequence, it was proposed that the initial formation of the Badain Jaran Desert began at 1.1 Ma. This time is agreed with the time proposed by Wang et al. (2015). Research on the lithofacies and magnetostratigraphy in the central Tengger Desert indicated that the initial formation of the Tengger Desert started at approximately 0.9 Ma (Li et al., 2014). This age also agrees with the time of river system formation in the eastern Qilian Shan (Pan et al., 2003; Pan et al., 2007), where the rivers flowed into the downstream of the Tengger Desert. Therefore, although the aridification in the desert area would provide the prerequisite conditions, the formation of large deserts in NW China required abundant sand materials which were transported by rivers from the uplifted regions as the Qilian Shan. The formation of large inland rivers might determine the formation time of the deserts in NW China.

Increase in the mass accumulation rate (MAR) in the Loess Plateau after 1.1 Ma were previously attributed to changes in aridity and the ranges of dust sources (Sun et al., 2008). In this text, we argue that the formation of large inland rivers, such as the Heihe River and the Shiyang River in NW China, played an important role in the evolution of

the Loess Plateau. Sediments trapped in the Alashan arid land is one of the major aeolian provenances of the Loess Plateau (Che et al., 2013) and beyond (Yang et al., 2007). Formation of the large inland rivers would greatly increase sediment supply from the NE Tibetan Plateau to the Alashan arid land. Combined with results of previous studies (Stevens et al., 2013; Nie et al., 2015), it is suggested that the formation of large inland rivers combined with incision of the Yellow River and its tributary channel systems in the Chinese Loess Plateau (Pan et al., 2007; Gao et al., 2008), which would greatly increase sediment supply to the dust sources, favored increased MAR in the Loess Plateau.

6. Conclusions

Sedimentary stratigraphy indicated that the sedimentary environment in the northern Jiudong Basin (west of the Hexi Corridor) significantly changed from a lake delta system to a large river system. The change occurred at ~1.1 Ma according to the magnetostratigraphy study. Before 1.1 Ma and back to at least 1.7 Ma, the alluvial system in the Jiudong Basin consisted of alluvial fans, transverse rivers, deltas and lakes. The significant change of the sedimentary environment 1.1 Ma is explained by the formation of the integrated Heihe River. The integrated Heihe River began to drain the Jiudong Basin and flowed into the Ejina Basin at ~1.1 Ma. Most of the terminal lakes in the Hexi Corridor disappeared, and independent transverse rivers became tributaries of the Heihe River. Moreover, the integrated Heihe River built a large fluvial fan in the Ejina Basin, which constrained the formation of the Badain Jaran Desert. The formation of the integrated Heihe River, together with other large rivers in north China, had a significant

impact on the evolution of the Loess Plateau. Drainage evolution may be a mechanism of the aeolian system evolution, alternative to climatic episodicity.

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