Magnetostratigraphy and paleoenvironmental events recorded in a late Cenozoic sedimentary succession in Huaibei Plain, East China

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ABSTRACT

The thick late Cenozoic deposits of Huaibei Plain provide a record of paleoenvironmental changes in the northern and southern transitional zone of eastern China, and of the evolution of the Asian monsoon system. A detailed magnetostratigraphic study of the Huainan (HN, 481.45-m deep) drill core from the center of Huaibei Plain reveals a magnetic polarity sequence from chron C4n.2n to chron C1n, spanning the interval from ~8 Ma to the present; it is the first reliable magnetostratigraphic chronology for the late Miocene to present for Huaibei Plain. The stratigraphic sequence contains three intervals characterized by major peaks in magnetic susceptibility (at depths of 58.8 m, 108.0 m and 312.4 m), which can be used as isochronous marker beds for regional stratigraphic correlations in Huaibei Plain. Major changes in sediment grain-size occur at 7.0 and 1.7 Ma, corresponding to pronounced environmental shifts, which we consider reflect the respective strengthening of the South Asian summer monsoon and East Asian winter monsoon at these times.

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1. Introduction

Reconstructing the Cenozoic environmental history of Huaibei Plain in eastern China is important for improving our understanding of the environmental evolution of the region, as well as of the East Asian monsoon (Jin, 1990; Shu, 2012; Guan et al., 2016; Zhang et al., 2016; Zhou et al., 2016). Most records of late Cenozoic changes in the East Asian monsoon are from terrestrial sedimentary sequences in central-western China (An et al., 2001; Ding et al., 2001; Guo et al., 2002; Sun et al., 2006; Deng et al., 2018) and from marine sediments in the South China Sea (Wang et al., 2003, 2016; Tian et al., 2011). In contrast, in eastern China, there are few terrestrial records with detailed chronologies, although such records may contribute substantially to our understanding of the evolution of the East Asian monsoon during the late Cenozoic.

Huaibei Plain contains sedimentary sequences that record changes in the East Asian monsoon during the late Cenozoic (Jin et al., 1987; Zhang, 2009; Qin et al., 2015; Zhou et al., 2016). To date, however, few studies of sedimentary sequences in Huaibei Plain have yielded long-term chronologies, since most of the investigations were made before 1990 when drilling did not reach bedrock, and therefore the recovered cores did not span the entire sedimentary succession. Preliminary paleomagnetic studies of Cenozoic sediments from Huaibei Plain were conducted more than two decades ago by Quan Jin and colleagues (Jin et al., 1987, 1988; Jin, 1990; Zhang, 2016) and by Zhenjiang Yu and colleagues (Yu, 1988; Yu and Huang, 1993; Yu and Peng, 2008). Jin (1990)
estimated that the base of the Huaibei Plain sedimentary sequence is older than 3.4 Ma, whereas Yu and Huang (1996) proposed an age greater than 4.5 Ma. However, these estimates were based on cores that did not reach bedrock and therefore they did not sample the base of the sedimentary sequence.

In 2012 a drilling project was conducted at Huainan in the center of Huaibei Plain with the aims of providing the first detailed long-term chronological framework for the region, and producing a detailed paleoenvironmental record and exploring its relationship with the evolution of the East Asian monsoon. During the project, a continuous 481.45-m-long drill core was obtained, consisting mainly of fluvo-lacustrine sediments, which were anticipated to provide information about late Cenozoic paleoenvironmental and paleoclimatic changes in eastern China and to document the evolution of the East Asian monsoon.

Here, we present the results of a detailed magnetostratigraphic and paleoenvironmental study of the entire fluvo-lacustrine succession. The magnetostratigraphy, together with age constraints provided by tephras layers and mammalian fossils, is used to interpret the observed changes in sediment grain-size and magnetic susceptibility. Finally, we discuss the results in the context of late Cenozoic environmental events in the region and their linkage with the evolution of the East Asian monsoon. Our results provide the first long-term chronological framework for sediments in Huaibei Plain, and they potentially improve our understanding of late Cenozoic environmental change in eastern Asia.

2. Regional environmental background, sedimentary lithology and sampling

2.1. Geological setting

Huaibei Plain, in central East China, is an alluvial–proluvial plain that developed at the base of a faulted basin during the Mesozoic, between the Yellow River and the Huaihe River. It covers an area of 37,421 km² (Hu et al., 2017) (Fig. 1a and b) and extends from 32°25′–34°35′N, 114°55′–118°10′E (Jin, 1990; Hu et al., 2014). It is bounded by the Yellow Sea to the east, the Funiu Mountains to the west, the Yellow River to the north, and the Dabie Mountains to the south (Cao, 2009). The surface of the plain is slightly inclined from northwest to southeast with a slope gradient of 1:8000, and the elevation above sea level varies from 15 to 50 m (Fig. 1b) (Jin, 1990; Wu et al., 2009). The 340-km-long Huaihe River flows from west to east through the southern part of Huaibei Plain, dividing it into northern and southern parts (Fig. 1b). The region lies within the warm and semi-humid climatic zone and has four distinct seasons (Jin, 1990).

Successions of unconsolidated Cenozoic sediments with thicknesses of several 100s of meters to more than 1000 m have accumulated in Huaibei Plain, with much thicker sequences in the west than in the east (Xie et al., 2013). These sediments consist primarily of alternating layers of sands, silty sands, and clays (Wu and Wu, 2014; Hu et al., 2017). The area has experienced tectonic subsidence during the Quaternary (Zhou et al., 2016) and most of the surface of the plain is covered by Quaternary strata, except for small areas of exposed bedrock in the northeast and west (Qian et al., 2015). Due to the flooding and diversion of the Yellow River, Holocene deposits in the northern part of the plain occur mainly along the tributaries of the Huaihe River, with typical thicknesses of several meters to ~10 m (Wu and Wu, 2014; Qian et al., 2015).

2.2. Lithology of the Huainan (HN) core

During April and May of 2012, a core (HN) was recovered using rotary drilling by the National Engineering Laboratory for Ecological Environmental Protection in Coal Mines. The core site is near the town of Guqiao in the middle of Huaibei Plain, about 30 km north of the Huaihe River (32°50.123′N, 116°30.167′E) (Fig. 1b). Drilling reached a depth of 481.45 m, with an average core recovery of 94%, and penetrated the entire sequence of Cenozoic sediments, reaching the underlying Permian sandstone bedrock. The core site is in the Huaibei subsidence zone, which has experienced continuous sediment accumulation during the late Cenozoic (Guo et al., 1992). There are no significant sediment discontinuities in the HN borehole and the sedimentary sequence is essentially complete, without significant hiatuses.

The lithology of the HN core was divided into eight lithological units (Units 1–8) from top to bottom, according to variations in grain size, bedding, color, and structure (Fig. 2). They are described as follows.

Unit 1 (0–52.4 m) is dominated by yellow clays, silty clays, and silts, and contains a few layers of calcium carbonate nodules. Layers of brown silty clays (21.2–29.0 m) and yellow silts (42.5–48.4 m) exhibit conspicuous horizontal bedding. The unit is interpreted as alluvial plain facies.

Unit 2 (52.4–69.6 m) is dominated by black to dull grayish-black clayey silts and silts with conspicuous horizontal bedding; it is interpreted as lacustrine facies.

Unit 3 (69.6–108.0 m) is dominated by yellow–yellowish silty clays and clays, containing numerous calcium carbonate nodules and a few layers of Fe–Mn concretions. Layers of yellow silty clays (79.6–86.4 m) and light grayish-black silty clays (98.4–103.6 m) exhibit conspicuous horizontal bedding. These characteristics indicate alluvial plain facies, like Unit 1.

Unit 4 (108.0–142.0 m) consists of yellowish-green or grayish-green silts, fine sands, and coarse sands with numerous calcium carbonate nodules and Fe–Mn concretions. A light grayish-green coarse sand layer, containing gravels with clasts of 0.5–2.0 cm diameter, is present from 132 to 136 m. The unit is interpreted as fluvial facies.

Unit 5 (142.0–207.2 m) consists of greenish-gray coarse sands, light grayish-gray and gray fine sands; clayey silts with conspicuous horizontal bedding occur from 166.4 to 175.6 m and from 184.0 to 188.0 m. The unit is interpreted as predominantly fluvial facies.

Unit 6 (207.2–398.4 m) is composed of dull greenish-gray coarse sand interbedded with greenish-gray sandy clays. The unit is interpreted as representing the alternation of fluvial and lacustrine facies.

Unit 7 (398.4–460.8 m) is dominated by greenish-gray clays with two layers of reddish silty clays (from 415.0 to 418.0 m and 431.2–432.4 m). The unit is interpreted as lacustrine facies.

Unit 8 (460.8–481.45 m) consists of dull reddish-silt and clayey silts, containing numerous calcium carbonate nodules and Fe–Mn concretions. The unit is interpreted as representing a predominantly terrestrial, oxidizing environment.

2.3. Sampling

A total of 451 samples for paleomagnetic analysis were obtained at a ~1 m interval throughout the core and placed in nonmagnetic square plastic boxes of 8 cm³ volume. The horizontal orientation of the samples is arbitrary because the core was obtained using mud-flush rotary drilling techniques and it did not have a consistent horizontal orientation. The samples were mainly taken from fine-grained intervals (clay, silt and fine-grained sand); however, several samples of coarser sediments were also obtained.
3. Methods

3.1. Grain-size measurements

A total of 1150 powder samples at ~40-cm intervals were obtained for measurements of grain-size (GS). Sample pretreatment consisted of the following. About 0.5 g of sample was weighed and placed in a 500-ml glass beaker. Then, 10 ml of 30% H2O2 was added to remove organic matter, followed by 10 ml of 10% HCl to remove carbonate, after heating to 140 °C. The residue was then dispersed with 10 ml of 0.5 N Na(PO3)6 and ultrasonicated prior to grain-size measurements. Grain size was measured using a Malvern Mastersizer 3000 laser grain-size analyzer with a measurement range of 0.01 μm to 3500 μm and 1% reproducibility. GS measurements were repeated three times and an average was taken. The measurements were made in the Key Laboratory of Cenozoic Geology and Environment, Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing, China.

3.2. Rock magnetic measurements

3.2.1. Magnetic susceptibility (χ)

A total of 2110 samples were collected at ~20-cm intervals for magnetic susceptibility measurements. χ was measured on 10 g of disaggregated sediment using a Bartington Instruments MS2 susceptibility meter at an operating frequency of 0.47 kHz. The measurements were repeated three times and an average was taken. The measurements were made in the Key Laboratory of Cenozoic Geology and Environment, Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing, China.

3.2.2. Temperature-dependence of magnetic susceptibility (χ–T curves)

Four samples were selected for measurements of the temperature-dependence of magnetic susceptibility (χ–T curves). The χ–T curves were measured using a MFK1–FA Kappabridge with a CS-3 temperature control system (Agico Ltd. Brno) in an argon atmosphere. The temperature was increased from room temperature to 700 °C with a ramping rate of 9 °C/min. Minor contributions to the magnetic susceptibility readings from the sample holder and thermocouple were subtracted from the final results.

3.2.3. Hysteresis loops and IRM acquisition

Four selected samples were used for measurements of hysteresis loops and isothermal remanent magnetization (IRM) acquisition. Hysteresis loops and IRM acquisition were measured using a MicroMag 3900 vibrating sample magnetometer (VSM, sensitivity = 0.5 × 10−9 Am²; Princeton Measurements Corp., U.S.A.) up to a maximum field of 1.5 T. IRM acquisition curves were unmixed using the methods and software of Kruiver et al. (2001) and were fitted with three components.

3.2.4. First-order reversal curves (FORCs)

The four selected samples were also used for measurements of first-order reversal curves (FORCs). FORCs were measured using a MicroMag 3900 VSM up to a maximum field of 1.5 T. FORC diagrams were generated using the FORCinel software package developed by Harrison and Feinberg (2013).

3.2.5. Paleomagnetic measurements

Stepwise alternating-field demagnetization (AFD) was conducted on 451 discrete samples using a 2G Enterprises Model 760-R cryogenic magnetometer installed in a field-free (<300 nT) space. The AF steps ranged from 2.5 to 20 mT, up to a maximum of 100 mT. All the above magnetic measurements were performed in Paleomagnetism and Geochronology Laboratory, Institute of Geology and Geophysics, Chinese Academy of Sciences (CAS), Beijing, China.
3.3. Geochemical analysis

The major element composition of Fe–Ti oxides from layers with high $\chi$ values (at 58.8 m, 108.0 m and 312.4 m) was measured using a JEOI JXA 8100 electron probe microanalyzer (EPMA, with a wavelength dispersive spectrometer (WDS)) at the State Key Laboratory of Lithospheric Evolution of the Institute of Geology and Geophysics, Chinese Academy of Sciences. The composition of 10 major elements (Na, Si, Mn, K, Mg, P, Fe, Ca, Al and Ti) of the Fe–Ti oxides was analyzed with a focused beam, an accelerating voltage of 15 kV, a beam current of 6 nA, and a beam diameter of 10 μm. Natural and synthetic oxides and silicate minerals of known composition were used as standards. Approximately 20–60 spots on the Fe–Ti oxides from each high $\chi$ layer were analyzed.

4. Results

4.1. Grain size

The lithology of the HN core is illustrated in Fig. 2, together with depth profiles of mean grain-size and magnetic susceptibility. The sequence consists mainly of silts and clays with occasional interbeds of sands (Fig. 2a). The mean grain size is 85 μm (Fig. 2b), and the average proportions of sands, silts, and clays are 34%, 32%, and
4.2. Rock magnetic properties

4.2.1. Magnetic susceptibility ($\chi$)

The $\chi$ values of the HN core are generally low (Fig. 2c); however, three abrupt increases in $\chi$ occur at the depths of about 58.8 m (layer A), 108.0 m (layer B), and 312.4 m (layer C) (Fig. 2c). The $\chi$ values at these depths range from $80 \times 10^{-8}$ m$^3$/kg to $300 \times 10^{-8}$ m$^3$/kg, about 10 times higher than elsewhere in the core. Overall, the $\chi$ of the coarse-grained fraction is low, and that of the fine fraction is high (Fig. 2d).

4.2.2. $\chi$–$T$ curves

$\chi$–$T$ curves provide useful information about changes in magnetic mineral composition during thermal treatment (Deng et al., 2001; Liu et al., 2014). $\chi$–$T$ curves were obtained for representative samples from the HN core and three types can be recognized: (i) This type is characterized by a major decrease in $\chi$ at about 585 °C, the Curie point of stoichiometric magnetite (Dunlop and Ozdemir, 1997), and by the cooling curves having much higher $\chi$ values than the heating curves, after cooling below 585 °C (Fig. 3a1, a4). This type also exhibits a major decrease in $\chi$ at ~585 °C, indicating magnetite, and a slight decrease is also observed at 675 °C (Fig. 3a2), which indicates the presence of hematite. The cooling curve is slightly higher than the heating curve after cooling below 520 °C. (iii) The heating curve of this type is the same as that of type (ii) (Fig. 3a3); however, the cooling curve is lower than the heating curve above ~275 °C, but much higher below ~275 °C.

In summary, all the heating curves exhibit a dramatic decrease in $\chi$ to near zero at 585 °C (Fig. 3a1–a4), which corresponds to the Curie temperature of magnetite; the heating curves also indicate that some of the samples also contain hematite. All the cooling curves indicate the neoformation of strong ferrimagnetic phases.
4.2.3. Hysteresis loops and IRM acquisition

Hysteresis loops and IRM acquisition curves are useful for characterizing the magnetic mineralogy and grain-size distribution of samples (Day et al., 1977; Dunlop, 2002). Four representative samples were analyzed. In two of the samples, the hysteresis loops are closed above 200–300 mT (Fig. 3b1, b2, b3), indicating the dominance of low-coercivity magnetic minerals. In the other samples, the loops are not closed at 400 mT (Fig. 3b4), indicating the presence of high-coercivity components, such as partially oxidized coarse magnetite grains and/or antiferromagnetic hematite (Deng et al., 2006). To evaluate this interpretation, IRM acquisition curves were unmixed following the procedure of Kruiver et al. (2001) and were fitted with three components. Three main
components were characterized in the samples: the main component has a coercivity \((B_c)\) range of 30–50 mT (component 2), while two additional minor components have \(B_c\) ranges of around 5–10 mT (component 1) and 300–600 mT (component 3). Two types of IRM acquisition curve are recognized: In the first type, component 2 represents about 90% of the SIRM, and components 1 and component 3 each represent about 5% (Fig. 3c1, c2). In the second type, component 2 represents about 80% of the SIRM, and components 1 and 3 respectively represent about 5% and 15% (Fig. 3c3, c4). These observations suggest that magnetite is the predominant magnetic mineral in the HN core sediments, while a small amount of hematite is also present.

### 4.2.4. FORC diagrams

FORC diagrams are very effective for characterizing the coercivity spectra and magnetic interaction fields of assemblages of magnetic grains (Roberts et al., 2000, 2014). The FORC diagrams for samples HN72 and HN107 (Fig. 3d1, d2) indicate pseudo-single domain (PSD) distributions, with concentric contours and a substantial vertical spread. In contrast, the FORC diagrams for the other samples indicate the dominance of single domain (SD) grains (Fig. 3d3, d4).

The same core samples’ T curves, hysteresis loops, IRM acquisition curves and FORC diagrams all confirm that the magnetic mineral assemblages of the sediments are dominated by PSD and SD magnetite, and that hematite is also present.

### 4.2.5. Paleomagnetic results

For most of the samples, at least 60% of the natural remanent magnetization (NRM) was removed after AFD at 80 mT, except for some samples containing hematite. Most of the samples yielded a stable characteristic remanent magnetization (ChRM) component after stepwise AFD up to 80 mT (Fig. 4), which indicates that magnetite is the dominant carrier of the ChRM.

ChRM directions were determined using at least four successive demagnetization steps >20 mT that define a linear trend towards the origin in orthogonal vector demagnetization plots. Samples with a maximum angular deviation (MAD) > 15° were excluded. The demagnetization results were evaluated using orthogonal plots (Zijderveld, 1967) drawn with PaleoMag software (Jones, 2002), and the principal component direction was calculated using a least-squares fitting technique (Kirschvink, 1980). Because of horizontal rotation of the HN core during drilling, the declination data are unusable and therefore the construction of the magnetic polarity sequence is based solely on inclination values.

A total of 279 samples showed unstable AFD behavior and were eliminated from the final data set, and the remaining 172 samples (38%) with reliable ChRM directions (average MAD of 8.2°) were used to establish the magnetic polarity sequence. About 110 unstable samples were from a coarse-grained interval (mainly medium and coarse sands) at the depth interval of about 110–310 m. This lithology may represent a sedimentary environment with ambient geomagnetic fields of assemblages of magnetic grains. All the 172 accepted ChRM directions of the HN core were used for a reversal test. In Fig. 5 the reversed polarities are inverted to their antipodes to test for a common mean for the normal and reversed magnetization directions. Furthermore, the HN core also passes the C quality reversal test (McFadden and McElhinny, 1990). Table 1 and Fig. 5 indicate that the obtained ChRMs likely represent a primary remanence (Tauxe, 1998).

The AFD results suggest that the HN core records 8 normal (N1–N8) and 7 reverse (R1–R7) polarity zones (Fig. 6). Each polarity zone was determined using at least three adjacent sampling levels.

### 4.3. Geochemical results

Three peaks in \(\gamma\) occur at depths of about 58.8 m, 108.0 m, and 312.4 m. According to the geochemical results, these layers of extremely high \(\gamma\) (Fig. 7) contain ilmenite, titanomagnetite and magnetite (Fig. 7d). In all three layers, FeO varies inversely with TiO₂, and the data display a linear trend. Fe–Ti oxides, especially titanomagnetite, are widely used as fingerprints to identify tephra layers in sediments (Shane, 1998, 2000; Jensen et al., 2008; Preece et al., 2011; Turner et al., 2011; Wang et al., 2014; Sun et al., 2016). Moreover, the presence of tephra can result in high \(\gamma\) values (Oldfield and Thompson, 1980; Hallett et al., 2001; Gehrels et al., 2008), and the high \(\gamma\) values and presence of titanomagnetite indicate that the three layers may contain tephra.

### 5. Discussion

#### 5.1. Correlation of the recognized magnetostratigraphy to the geomagnetic polarity timescale (GPTS)

Previous studies have reported two mammalian fossil-bearing layers (at depths of 46.2 m and 78.9–82.85 m) in a drill core from Panji, located about 15 km east of the HN core site (Fig. 1). The mammal fossils found at the depth of 46.2 m in the Panji core are Cervus grayi and Cervus unicolor; of Middle Pleistocene age (Jin, 1990). Stegodon orientalis fossils, of Early Pleistocene age, were discovered in the depth interval of 78.9–82.85 m in the same core (Jin, 1990). The Panji core site is very close to the HN core site and the strata are nearly horizontal in the area (Fang et al., 2010); thus, these two fossil-bearing layers can be used to constrain the
According to previous research, only three documented volcanic eruptions occurred in the Huaihe River drainage area during the late Cenozoic. Nvshan volcano, about 150 km east of the HN core site, erupted twice during the Quaternary, at 0.64 Ma and 1.41 Ma (Xia et al., 1994; Chen and Peng, 1988). In addition, Pingmingshan volcano, about 290 km east of the HN core site, erupted at ~6.1 Ma (Liu, 1992; Wang, 2011). Interestingly, there are three potential tephra layers (indicated by extremely high $c$ values and the presence of titanomagnetite) in the HN core, which can tentatively be correlated with the three known eruptions. Layer A (58.8 m) correlates with the Nvshan eruption at 0.64 Ma, layer B (108.0 m) with the Nvshan eruption at 1.41 Ma, and layer C (312.4 m) with the Pingmingshan eruption at 6.1 Ma. The three layers in which major peaks in $c$ occur (at depths at 58.8 m, 108.0 m and 312.4 m, dated to 0.64 Ma, 1.41 Ma and 6.1 Ma, respectively) can be used as isochronous marker beds for stratigraphic correlation between different late Cenozoic sedimentary sequences in Huaibei Plain.

These correlations provide useful constraints on the observed polarity sequence which facilitate its correlation with the geomagnetic polarity time scale (GPTS) (Gradstein et al., 2012; Ogg, 2012), as shown in Fig. 6. Layer C (312.8-m depth, 6.1 Ma) is within the upper part of polarity zone N7. Based on the age control points provided by the tephrostratigraphy, magnetozone N7 can be correlated to chrons C3An.1n to C3An.2n (because N7 is a long normal polarity zone); R7 can be correlated to chron C3An.2r; and N8 can be correlated to chrons C3Bn–C4n.2n, which are mainly normal polarity zones. For the upper part of N7, owing to the relatively coarser grain size of the stratigraphic interval at 217–304 m, the paleomagnetic results are of relatively low resolution. Nevertheless, magnetozones R5–R6 can potentially be correlated to chrons C2Ar–C3r, predominantly of reversed polarity (the Gilbert reverse chron), with one normal polarity zone (N6). Thus, magnetozone N5 can be correlated to chrons C2An.1n–C2An.3n, dominantly of normal polarity (the Gauss normal chron). Layer A (0.64 Ma) is at the depth of 58.8 m, within the upper part of magnetozone R1. Therefore, magnetozones N1 and N2 (0–72 m in the HN core) should correspond to the Brunhes normal chron (the Gauss normal chron). Layer A (0.64 Ma) is at the depth of 58.8 m, within the upper part of magnetozone R1. Therefore, magnetozone N1 and N2 (0–72 m in the HN core) should correspond to the Brunhes normal chron of the GPTS (Gradstein et al., 2012), which is consistent with the Cervus grayi and Cervus unicolor fossil age (46.2 m, Middle Pleistocene) from the Panji core (Jin, 1990). Within the Brunhes chron, there is a reversed polarity zone (R1) which probably corresponds to the Delta geomagnetic excursion at ca. 680 ka (Champion et al., 1988; Thouveny et al., 2008), given that layer A (640 ka) is within R1. Magnetozones R2–R4 should be correlated to the Matuyama reversed chron. Layer B (108 m depth, 1.41 Ma) is within the lower part of magnetozone R3, which can thus be correlated to chron C2r.2r. Within the Matuyama chron, there are two normal magentozones (N3 and N4) that may correspond to the C1r.1n (Jaramillo) and C2n (Olduvai) normal subchrons,
respectively (Gradstein et al., 2012; Heirtzler et al., 1968; Valencio et al., 1970). This correlation is consistent with the biochronology based on the occurrence of fossils of *Stegodon orientalis* (78.9–82.85 m, Early Pleistocene) in the Panji core (Jin, 1990). Finally, a robust magnetostratigraphic chronostratigraphic framework was constructed for the HN core with a thickness of 481.45 m (Fig. 6).

The age of the base of the HN core is estimated at ~8.0 Ma based on extrapolation of the sediment accumulation rate derived from the two lowermost polarity reversal boundaries. Thus, HN core spans the interval from about 8.0 Ma to the present. An age-depth model was determined via linear interpretation between the ages of the two boundaries. According to our age model, the Middle/Late Pleistocene, Early/Middle Pleistocene (Matuyama/Brunhes, M/B), Pliocene/Pleistocene (Gauss/Matuyama, Ga/M), Gilbert/Gauss (Gi/Ga), and Miocene/Pliocene boundaries correspond to depths of 11.6 m, 72.0 m, 136.0 m, 211.0 m, and 286.1 m, respectively.

5.2. Regional magnetostratigraphic correlation

Paleomagnetic records from Huaibei Plain have only been generated for seven cores (Linquan, Boxian, Woyang, Yingshang, Huangkou, Mengcheng, and Guzhen; Fig. 1) (Jin, 1990). The M/B boundary in these seven cores was at the depths of 89.00 m, 92.00 m, 93.00 m, 79.00 m, 72.00 m, 91.50 m, and 84.50 m, respectively (Fig. 8); and the respective depths of the Ga/M boundary were 142.64 m, 159.95 m, 141.50 m, 114.03 m, 127.38 m, 134.82 m, and 111.84 m (Fig. 8) (Jin, 1990). Thus, within Huaibei Plain, the M/B boundary occurs within the depth range of ~72–92 m, and the Ga/M boundary within the depth range of ~112–160 m. The M/B and Ga/M boundaries occur at the respective depths of 72 m and 136 m in the HN core (Fig. 8), and thus, they are consistent with the results from elsewhere in Huaibei Plain, supporting the reliability of our magnetostratigraphic framework.

Although paleomagnetic studies were carried out on these seven other cores, only two reached bedrock, Huangkou and Mengcheng (Fig. 8), from the margins of the Huaibei Plain (Fig. 1). Jin (1990) estimated the age of the basal Cenozoic sediments in the Huaihe Basin to exceed 3.4 Ma, and Yu and Huang (1996) proposed a basal age exceeding 4.5 Ma. However, these studies were not based on cores that reached bedrock in the center of the Huaibei Plain, and thus the proposed ages are unlikely to accurately represent the initiation of deposition in the basin. The HN core was recovered from a location close to the center of the Huaibei Plain and it did reach bedrock, meaning that its basal age may closely represent the beginning of sediment accumulation in Huaibei Plain. According to the magnetostratigraphy of the HN core, as determined in the present study, accumulation of Cenozoic sediments in Huaibei Plain began by at least 8.0 Ma.

The magnetostratigraphies for Linquan, Boxian, Woyang, Yingshang, Huangkou, Mengcheng and Guzhen are cited from Jin (1990).

5.3. Paleoenvironmental events in Huaibei Plain

Our results can be used to reconstruct the late Cenozoic...
environmental evolution of Huaibei Plain. Sediment grain-size has been used as a paleoenvironmental proxy in many studies, for example as a measure of the intensity of the East Asian winter monsoon in Chinese loess deposits, and as an index of hydrodynamic conditions in fluvio-lacustrine sediment sequences (Reading, 1978; Bagnold and Barndorff-Nielsen, 1980; An et al., 1991; Liu and Ding, 1993; Ding et al., 1994; Singh et al., 2007; Li et al., 2017). Generally, in aqueous environments, coarse particles...
indicate a high-energy hydrodynamic environment (i.e. fluvial), and fine particles may indicate a low energy hydrodynamic environment (i.e. lacustrine) (Sun et al., 2002; Singh et al., 2007; Li et al., 2017). Given the constraints of our age–depth model, the sediments aged from 8 to 7 Ma contain a high proportion of clay (average 43.1%) and silt (average 44.6%), but a relatively low proportion of sand (average 12.3%). However, in the sediments aged from 7.0 to 1.7 Ma the proportion of sand increases substantially (average 51.1%), whereas there are decreases in the proportions of clay (average 17.5%) and silt (average 31.4%). From 1.7 to 0.0 Ma, silt dominates the sediments (average 69.3%), whereas the sand content (average 9.2%) decreases substantially (Fig. 9). Thus, we interpret the grain size and sedimentary facies of the HN core to indicate a mainly lacustrine environment before 7.0 Ma, a shift to a fluvio-lacustrine environment from 7.0 to 1.7 Ma, and finally the development of a predominantly alluvial plain environment after 1.7 Ma.

In South Asian marine sediments, tests of the planktonic foraminifer Globigerina bulloides show a pronounced increase in abundance at ~7 Ma, which also corresponds to a shift from C3 to C4 terrestrial vegetation (Fig. 9a) (Quade et al., 1989). Marine records also show that the strength of the South Asian summer monsoon was at a maximum during the late Miocene (~7 Ma) (Gupta et al., 2015). At all sites drilled in the Bengal Fan during ODP Leg 116, a shift in clay mineralogy to a smectite–kaolinite assemblage, and a decrease in sediment accumulation rate, occurred at ~7 Ma, indicating a shift to a warmer and more humid climate (France-Lanord et al., 1993). Sr flux to the Bay of Bengal also decreased at ~7 Ma, caused by decreasing erosion rates in the Himalaya region (Derry and France-Lanord, 1996). These changes suggest that the South Asian monsoon began to strengthen at 7 Ma. In northern China, deposition of the Lantian red clay began at ~7 Ma, which may be related to changes in the East Asian monsoon (Kaakinen, 2005). In the North Pacific, eolian dust records show that dust flux gradually decreased between ~7.7 Ma and ~7.0 Ma, reached a minimum at ~7 Ma, and gradually increased thereafter (Fig. 9b) (Rea et al., 1998). This evidence indicates an important change in the climate of the Asian interior at ~7 Ma. Thus, several significant climatic changes (involving the South Asian monsoon, the East Asian monsoon and the Asian interior) occurred at ~7 Ma, which is concordant with the shift in grain size observed in the HN core (Fig. 9c).

In addition, records of glacial till, glaciomarine diamictite, and ice-rafted detritus suggest that substantial glaciation began in Greenland as early as ~7 Ma (Larsen et al., 1994), which may have resulted in the weakening of the East Asian summer monsoon and the strengthening of the winter monsoon.

The strengthening of the South Asian monsoon, weakening of the East Asian summer monsoon and strengthening of the winter monsoon imply that a major adjustment of the Asian monsoon system occurred at ~7 Ma. The maximum strength of the South Asian summer monsoon was at ~7 Ma (Gupta et al., 2015). Modern meteorological data show that an enhanced South Asian monsoon transports large amounts of water vapor to Huaibei Plain, which

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Fig. 9. Comparison of late Cenozoic paleoenvironmental records from different regions. (a) δ¹³C (PDB) record from Hole 807A, Ocean Drilling Program (ODP) Leg 130, on the northern rim of the Ontong Java Plateau; the record indicates a change from C3 to C4 plant dominance in South Asia (Quade et al., 1989). (b) Eolian dust flux record from ODP Sites 885/886 in the North Pacific (Rea et al., 1998). (c) Grain size variations in the HN core, East China (this study). (d) Smoothed (50-pt running mean) record of benthic foraminiferal δ¹³C data from ODP Site 607 in the North Atlantic (Hodell and Venz-Curtis, 2006). (e) Estimated equatorial zonal sea-surface temperature (SST) gradient between 159°E and 95°W (Wara et al., 2005). (f) Difference in δ¹³C values between O. universa and G. bulloides (two species of planktonic foraminifera that proliferate in different seasons) at ODP Site 1014 (eastern Pacific), which are thought to be caused by changes in seasonality (red dots) (Ravelo et al., 2004). (g) Record of mass accumulation rates (MAR) of biogenic carbonate at ODP Site 1014 (Ravelo et al., 2004). (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)
increases the rainfall (Mao et al., 2009); this in turn strengthens the fluvial dynamics resulting in a coarsening of the sediment grain-size. Therefore, we infer that the change from a lacustrine environment (fine grain-size) to a fluvial environment (coarse grain-size) in Huaibei Plain at ~7.0 Ma was directly linked to the adjustment of the Asian monsoon system. A rapid decrease in the δ13C value of benthic foraminifera at ODP Site 607 in the North Atlantic happened at ~1.7 Ma (Fig. 5d) (Hodell and Venz-Curtis, 2006). In addition, based on δ13C records, temperatures in the Western Equatorial Pacific (WEP) increased by ~2 °C at ~1.7 Ma (Fig. 3e) (Wara et al., 2005). A dramatic decrease in the calcite mass accumulation rate (CaCO3-MAR) at ODP Site 1014 in the eastern Pacific also occurred at ~1.7 Ma, most likely because of an increase in seasonality (Fig. 9f, g) (Ravelo et al., 2004). These records indicate that the global climate, including in the Pacific, experienced a major shift at ~1.7 Ma. Dramatic changes in the climate of the Pacific may indicate a major shift in the ocean-atmosphere system of the Pacific Ocean - Asian continent that would have substantially affected the East Asian monsoon system (An, 2000; Ding and Chan, 2005; Chu et al., 2017), which in turn would have impacted the climate and environment of Huaibei Plain in East China.

6. Conclusions

We have conducted a paleomagnetic, rock magnetic, geochemical and grain size study of a 481.45-m-long HN core, recovered from the central part of Huaibei Plain in eastern China. The magnetic mineralogy of the sediments is dominated by magnetite, with hematite also present. The magnetostratigraphic record spans chron C4n.2n to C1n of the GPTS and the data suggest a magnetostratigraphic age of ~8 Ma for the base of the sedimentary sequence in central Huaibei Plain, which represents a minimum age for the initiation of the accumulation of Cenozoic sediments in the region. The Matuyama/Brunhes, Gauss/Matuyama, Gilbert/Gauss, and Miocene/Pliocene boundaries are located at the depths of 72.0 m, 136.0 m, 211.0 m, and 286.1 m, respectively. In addition, the sequence contains three layers with major peaks in magnetic susceptibility, at the depths of 72.0 m, 136.0 m, 211.0 m, and 286.1 m, respectively. These peaks can be used as isochronous markers for stratigraphic correlation between different late Cenozoic sedimentary sequences of Huaibei Plain. Pronounced environmental shifts in Huaibei Plain at ~7 and ~1.7 Ma are considered to reflect the respective strengthening of the South Asian summer monsoon and the East Asian winter monsoon.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.quascirev.2018.09.041.