Catastrophic drought in East Asian Monsoon region during

Heinrich event 1

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Abstract: Heinrich event 1 (H1) is an important millennial climate event during the last deglaciation. The substantial decreasing of monsoon strength in the East Asian monsoon region during the H1, as shown by stalagmite $\delta^{18}\text{O}$ records, has been attributed to the southward shift of the intertropical convergence zone (ITCZ), which is caused by the slowdown/collapse of the Atlantic meridional overturning circulation (AMOC). However, records from different Asian monsoon regions show various trends in precipitation changes during the H1, and these trends cannot be solely interpreted by the southward shift of the ITCZ. In the present study, we reconstructed time-series of East Asian monsoon precipitation between 25,000 and 10,000 a BP from floodplain sediments in the Huai River Basin. A white sediment layer, distinct from other layers in the profile, contains significantly low TOC, tree pollen and fern spore contents, and more positive $\delta^{13}\text{C}_{\text{org}}$, and it is deposited during the H1 event. The determined TOC, pollen and $\delta^{13}\text{C}_{\text{org}}$ time-series, together with previously reported stalagmite $\delta^{18}\text{O}$, indicate a catastrophic (severe) drought in Jianghuai Region, one of the East Asian monsoon regions, during the H1. The La Niña condition in tropical Pacific likely also contributes to the catastrophic drought in Jianghuai Region and the precipitation variations in the Asian monsoon region during the H1.

Keywords: Heinrich event 1; Asian summer monsoon; ENSO; catastrophic drought; floodplain sediment

1. Introduction

A series of climate events, including Heinrich events and Dansgaard-Oeschger cycle (Heinrich, 1988; Bond et al., 1993; Dansgaard et al., 1993), occurred during the last glaciation, indicating an unstable climate in glacial periods. One of the important events, the H1 event, occurred between ~15,000 and 18,000 years before present, was recorded in Greenland ice cores (Bond et al., 1993), deep sea sediments (Bond et al., 1993), loess (Porter and An, 1995), stalagmite (Wang et al., 2001) and lacustrine sediments (Stager et al., 2011) etc, and was proposed to be a global climate event.

The main forcing of the H1 has been generally attributed to the abrupt input of cold and fresh water into North Atlantic Ocean, which caused the slowdown or even collapse of the Atlantic meridional overturning circulation (AMOC) (McManus et al., 2004; Denton et al., 2010). The H1 event recorded in monsoon regions has been considered to be caused by the AMOC-forced southward movement of the intertropical convergence zone (ITCZ) (Thomas et al., 2009; Denton et al., 2010). During the H1, stalagmite $\delta^{18}\text{O}$ became more positive in the East Asian monsoon region (Wang et al., 2001), precipitation decreased (Stager et al., 2011), and dust flux increased (Peck et al., 2004) significantly in Africa. Overall, Afro-Asian monsoon region became dry.

The slowest AMOC during the H1 was shown by both proxies (McManus et al., 2004; Shakun et al., 2012; Lynch-Stieglitz et al., 2014) and model results (Ritz et al., 2013). If the H1 event in monsoon regions was only forced by the southward shift of ITCZ, the H1 in northern hemisphere monsoon regions would be the driest period during the past 25,000 years, as supported by stalagmite $\delta^{18}\text{O}$ in southern China (Wang et al., 2001).

However, monsoon precipitation proxies in other regions, such as stalagmite $\delta^{18}\text{O}$ in Indian monsoon region (Shakun et al., 2007) and ^{10}Be concentrations in loess from Loess Plateau (Zhou et al., 2012), showed relatively less reduction of precipitation during H1 and even weakest summer trade wind during the Last Glacial Maximum (LGM). Thus the catastrophic drought in Afro-Asian monsoon region during the H1 was not only caused by the southward movement of ITCZ, but also affected by other factors such as sea surface temperature (SST) (Stager et al., 2011). And it's essential to reconstruct more records to further study the forcing mechanisms of the H1 in the Asian monsoon region.

Huai River Basin is located in the East Asian monsoon region and modern north boundary of the subtropics; thus it is sensitive to both ITCZ shift and temperature change. In the present study, we reconstructed time-series of climate change at a study site in the Huai River Basin during the LGM and the last deglacial period, focused on climate changes during the H1, and examined the forcing mechanisms of the H1 in the Asian monsoon region.

2. Materials and methods

Our study site, labeled as LW, is located on the floodplain of Huai River in northern Anhui Province, China. The upper part of the site is an archaeological site, which has been suggested to be a burial area of about 5,000 years ago. The profile is about 400 meters away from the Qianhe River, a branch of the Huai River (Fig. 1). Nowadays, plantations are on the top of the profile and farmland is near the study site. Because the upper part of the profile is archaeological deposits, the top of the profile is 1-2 meters

higher than neighboring farmland.

The LW profile was 4.35 m long from the top to the bottom where groundwater appeared, and the sampling interval was 1 cm. All the 435 samples were analyzed for total organic carbon (TOC), total nitrogen (TN), Na element concentrations, and grain size. 24 samples were analyzed for $\delta^{13}\text{C}_{\text{org}}$, 22 samples were analyzed for pollen concentrations, and 10 samples were chosen for AMS ^{14}C dating.

About 0.2 g of each sample was weighed, 10-20 ml H_2O_2 solution (30%) was added, and the mixture was heated to 100 °C for 0.5 h to remove organic matter and bathed in 10 ml HCl solution (10%) for 48 h to remove calcareous cement and shell materials. All samples were fully desalted and dispersed by adding 10 ml $(\text{NaPO}_3)_6$ (10%) and by ultrasonic treatment for 10 minutes before measurement. The grain size was analyzed using the LS13320 laser diffraction particle size analyzer (Beckman Coulter, Inc.) at the Institute of Polar Environment, University of Science and Technology of China. The analysis range of the analyzer was 0.04~2000 μm , the size resolution was 0.01 Φ , and the relative error of repeat test was less than 2%.

The chemical volumetric method was used to measure TOC content with a duplicate error of 5% (Loring and Rantala, 1992) at the Institute of Polar Environment, University of Science and Technology of China. The contents of Na_2O were measured using X-Ray Fluorescence Spectrometry of the Institute of Geology and Geophysics, Chinese Academy of Science. TN content was measured by Vario EL III (Elementar Company) with an error smaller than 1% at the Institute of Polar Environment, University of Science and Technology of China.

$\delta^{13}\text{C}_{\text{org}}$ were measured using a Finnigan MAT 251 gas source mass spectrometer at

the Institute of Earth Environmental, Chinese Academy of Sciences. Isotopic ratios are reported using δ -notation as per mil (‰) differences relative the V-PDB standard for C. The overall precision of the $\delta^{13}\text{C}$ analyses was less than 0.2‰.

Pollen samples were processed using heavy liquid separation (Moore and Webb, 1978) and acetolysis (Erdtman, 1960). Pollen analysis was performed on 22 samples (100-200g each) at the Institute of Vertebrate Paleontology and Paleoanthropology, Chinese Academy of Sciences. Pollen percentages are calculated from the total pollen count of arboreal and non-arboreal taxa identified in each pollen spectrum.

AMS ^{14}C dating was carried out using TOC at the Center for Applied Isotope Studies at the University of Georgia. The result was calibrated to calendar age by the computer program IntCal04 (Reimer et al., 2004). All the dating results were in calendar age in this study.

3. Results and discussion

3.1. Stratigraphy, lithology and time-series of proxies

As observed in the field, the color of different layers in the LW profile has obvious changes (Fig. 2). Because the upper part (0-140 cm depth) contains a lot of burned red clay, the color is relatively deeper. Sediments at depth of 140-170 cm and 210-250 cm are grey yellow, and those below 270 cm are brown. Notably, sediments of 170-210 and 250-270 cm are white, and the white layer at 250-270 cm is more visible (Fig. 2). Because the burned red clay in the upper part indicates the influence of ancient human

activities, we will only reconstruct and discuss natural environmental changes using the sediments below 140 cm.

Median grain size of the sediments below 280 cm changes little with an average of 12.0 μm , indicating a relatively stable sedimentary environment during that period. However, median grain size of sediments between 140-280 cm, with an average of 14.6 μm and a range of 6.5-20.9 μm , varies greatly. Sediments of 170-200 cm, 230-240 cm and 250-280 cm are coarser.

TOC content of the sediments below 140 cm varies between 0.01-0.57%, with high values of ~0.50% in 140-170 cm and 210-250 cm and low average values of 0.19% in other layers. The lowest TOC content is about 0.01% around 260 cm. TN content below 140 cm has relatively little change, with an average of 0.044% (Fig. 2). Between 140 and 280 cm, the trend of Na content is similar to that of median grain size (Fig. 2).

Measured $\delta^{13}\text{C}_{\text{org}}$ range from -18.27‰ to -20.31‰ with an average of -19.28‰. The $\delta^{13}\text{C}_{\text{org}}$ are more positive at 190 and 270 cm and the most negative at 230 cm. Trend of the $\delta^{13}\text{C}_{\text{org}}$ is similar to that of TOC (Fig. 2).

Among the 22 samples, 14 samples contain no pollen grains. Other eight samples contain about 300 grains of pollens or fewer; thus we only used these eight pollen records to study vegetation change in the LW profile. A total of 27 taxa were obtained from the analysis of the eight samples. Pollen types were classified into tree and herb groups (Fig. 3). The main pollen types include *Pinus*, *Picea Cupressaceae*, *Artemisia*, *Chenopodiaceae*, *Poaceae*, *Asteraceae*, etc.

3.2. Chronology

Ten AMS ^{14}C ages of the LW profile are shown in Table 1. To investigate the “reservoir effect”, we compared ^{14}C ages of charcoal and bulk sediment at 173 cm. The two ^{14}C ages are consistent with each other and within the error range and the sedimentary phase of sediments used for ^{14}C dating does not change, suggesting that the “old carbon effect” is insignificant or not present. Because the sampling depths of 215 cm and 282 cm are close to the slope surface (Fig. 2), the two ^{14}C ages are possibly polluted by “new carbon” leached from covering soil of the profile. Contamination by even a few modern carbon materials could cause large shifting of ^{14}C age (Olsson, 1991). Thus we deleted the two ages and used other eight ages to determine the chronology of the profile (Fig. 4). The average sedimentation rate of the profile is $16.20 \text{ cm}\cdot\text{ka}^{-1}$, and the time period recorded by the core is about ~26,850 years.

3.3. Sedimentary phase

Because the LW profile is located in the Huai River floodplain, sediments are mostly floodplain deposition. The deposition process of the sediments is likely gradual and relatively continuous. Firstly, earlier studies (Jin, 1990; Yu and Peng, 2008) have shown that the Huai River Basin is located in a tectonic subsidence region in Quaternary. Secondly, the fine grain size of the sediments (Fig. 2) indicates that the sediments are not deposited under strong hydrodynamics. Using C-M diagram of grain size (Fig. 5) and grain size distributions of typical samples (Fig. 6), we further confirmed the sedimentary phase of the profile as floodplain deposition.

C-M diagram ([Passega, 1964](#)), which shows the relationship of median grain size and the coarsest 1% of grain size, has frequently been used to study sedimentary phase. The C-M diagram of sediments from the LW profile (Fig. 5) shows little change in M but remarkable one in C before 10,000 a BP, indicating a suspension transport environment.

The frequency and cumulative distribution curves of grain size (Fig. 6) of the LW profile are consistent with those of typical shallow lake sediments ([Sun et al., 2002](#)), but the frequency distribution curve of LW has more peaks. Thus the LW sediments were deposited under shallow water and might experience pedogenesis processes, further confirming the nature of floodplain deposition.

3.4. H1 event recorded in LW profile

In the LW profile, the age at depth of 250-270 cm is around 16,000 a BP, the occurrence time of the H1 event (around 16,800 a BP). Thus the white layer around 260 cm is likely from deposition during the H1. Element analysis shows that Na concentrations in sediments of the white layer are higher than those in other layers (Fig. 2). As shown by earlier studies ([Shan et al., 1986](#)), salt could be transported to the surface of the soil under evaporation during very dry periods. Thus the higher Na contents of the white layer in the LW profile likely indicate drier environment around 16,000 a BP. However, if the place had a water-logged environment during the H1, Na concentrations might also increase. Thus further studies are needed to test whether the white layer was deposited during extremely dry period.

Because Huai River Basin is located in the typical monsoon region – the East Asian

monsoon region, the changes in vegetation cover in this region is strongly influenced by the monsoon evolution. When summer monsoon strengthens, vegetation cover and terrestrial TOC transported by rivers would increase. Thus TOC in sediments can be used to study summer monsoon evolution (Xiao et al., 2006). As shown in Fig. 7, the monsoon proxies, as defined by TOC, are in excellent agreement with the stalagmite $\delta^{18}\text{O}$ of Hulu Cave (Wang et al., 2001), which has been commonly used as Asian summer monsoon proxy. Hulu Cave is located in the lower Yangtze River Basin, where moisture change might be caused by the East Asian monsoon; thus stalagmite $\delta^{18}\text{O}$ signal of the cave likely indicate East Asian monsoon variation, though climatic implication of stalagmite $\delta^{18}\text{O}$ in Chinese caves remains debatable, particularly during the Holocene (Yang et al., 2014; Caley et al., 2014; Chen et al., 2015; Liu et al., 2015). In addition, the consistent changes between TOC and stalagmite $\delta^{18}\text{O}$ provide further support that sediments of the LW profile could be relatively continuously deposited.

Time-series of TOC show relatively weak summer monsoon around the LGM. The weakest summer monsoon occurred around 16,000 a BP, and strengthened after ~15,000 a BP. Between 15,000 and 12,600 a BP, high TOC values indicate strong summer monsoon. The weak monsoon between 12,600 and 11,500 a BP is temporally consistent with the Younger Dryas (Hoek, 2008, Chen et al., 2015).

Notably, the weakest summer monsoon around 16,000 a BP during the H1 event, as recorded by TOC of sediments in the LW profile, is consistent with that recorded by stalagmite $\delta^{18}\text{O}$ (Fig. 7). Thus during the H1, precipitation in the Huai River Basin decreased substantially. The white layer, with high Na content, in the LW profile was formed under dry environment, and it provides a visible evidence for the H1 event in

stratigraphic profile.

$\delta^{13}\text{C}$ of vegetation has been extensively used for studying precipitation changes (Hong et al., 2003; Liu et al., 2011). Because decreased temperature would result in more negative $\delta^{13}\text{C}$ (Wang et al., 2002) of vegetations, the positive $\delta^{13}\text{C}$ could not be caused by lower temperatures during the H1 event (Alley, 2000). The more positive $\delta^{13}\text{C}_{\text{org}}$ in the LW profile at around 16,000 and 12,000 a BP (Fig. 8) might be attributed to the following two possible mechanisms. Firstly, $\delta^{13}\text{C}_{\text{org}}$ in the LW profile varied from -18.27‰ to -20.31‰, which are located between $\delta^{13}\text{C}$ values of C3 and C4 plant (Ehleringer et al., 1997). Thus more positive $\delta^{13}\text{C}$ values might be caused by increasing of C4 plants, which are more tolerant to drought. Secondly, spatial distributions of $\delta^{13}\text{C}$ and precipitation show that $\delta^{13}\text{C}$ of vegetation decreases as precipitation increases (Kohn, 2010). In consequence, $\delta^{13}\text{C}_{\text{org}}$ of sediments in the LW profile became more positive at around 16,000 and 12,000 a BP (Fig. 8), indicating relatively less precipitation during the two periods. Though pollen grains are not enough for statistical analysis of vegetation and environment change, tree pollen and fern spore concentrations can be used as a reference. The two records (Fig. 8) roughly show lowest concentrations around 16,000 a BP, supporting a dry climate during the H1 event.

3.5. Forcing mechanisms of catastrophic drought during the H1 event in the East Asian monsoon region

During the H1, it is generally believed that the AMOC slowed down or even collapsed, and the ITCZ moved southward, causing weak summer monsoon and drought

in the northern hemisphere monsoon region (Wang et al., 2001; Stager et al., 2011). If the ITCZ shift were the only factor, decreases in summer monsoon precipitation would be expected for the entire northern hemisphere monsoon region. Both TOC of sediments in the Huai River Basin reconstructed in the present study and stalagmite $\delta^{18}\text{O}$ record from Hulu Cave in the Yangtze River Basin (Wang et al., 2001) show catastrophic drought. Other records in loess also show dry conditions during the H1 event (Porter and An, 1995; Ren et al., 1996; Chen et al., 1997; Ding et al., 1998; Wang et al., 2012). However, ^{10}Be record in loess from northern China (Zhou et al., 2012) and stalagmite $\delta^{18}\text{O}$ record in Yemen (Shakun et al., 2007) show higher precipitation during the H1 than during the LGM (Fig. 9), and some records in the Indian monsoon region show weaker monsoon during the H1 than during the LGM (Chen et al., 2014). These discrepancies indicate that the regional precipitation variation in the Asian monsoon region is influenced by other factors as well. As suggested by Stager et al (2011), the catastrophic drought during the H1 in Africa is not only controlled by southward shift of the ITCZ, but also by the changes in sea surface temperature.

El Niño-Southern Oscillation (ENSO) has substantial influence on monsoon precipitation in the Asian monsoon region. As indicated by instrumental observations, precipitation generally increases in southern China but decreases in most of the Indian monsoon region during strong ENSO period (El Niño condition), and opposite changes occur during weak ENSO period (La Niña condition) (Zhang, 2001).

ENSO is usually characterized by SST difference between the eastern and the western tropical Pacific. The SST difference is larger under La Niña condition and smaller under El Niño condition. Our calculated ΔSST using earlier reconstructions of

SST in the eastern (Dubois et al., 2011) and the western tropical Pacific (Mohtadi et al., 2010) (Fig. 10) shows larger Δ SST during the H1. This La Niña condition would cause regional differences in precipitation, such as less precipitation in southern China. We thus propose that the precipitation changes in the Asian monsoon region are driven by both the southward shift of ITCZ and ENSO conditions. During the H1, the combined effect of the southward shift of ITCZ and weak ENSO conditions caused the regional differences in precipitation in Asian monsoon region, such as catastrophic drought in Jianghuai Region of China.

4. Conclusions

In the present study, we reconstructed time-series of monsoon precipitation between 25,000 and 10,000 a BP in the Huai River Basin. The white layer of high Na contents, deposited during the H1, provides a visible evidence for H1 in the field. Time-series of TOC, tree pollen percentage and $\delta^{13}\text{C}_{\text{org}}$ all show catastrophic drought in the Huai River Basin during the H1, in consistence with stalagmite $\delta^{18}\text{O}$ of Hulu Cave in southern China. However, records in northern China and some of the Indian monsoon region both show increased precipitation during the H1 relative to other periods of the Last Glacial Maximum. Therefore, we suggested that the precipitation changes during the H1 in the Asian monsoon region are likely caused by both the southward shift of ITCZ and the La Niña -like conditions.

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428

Figure captions

Fig. 1 Sampling site (Our study site is shown by red solid triangle, sites of earlier studies are shown by red solid circles). Base map is drawn by ArcGIS 10.0 software.

Fig. 2 Stratigraphy and proxies of the LW profile.

Fig. 3 Pollen concentrations in the LW profile.

Fig. 4 Profiles of AMS ^{14}C age for the LW profile.

Fig. 5 Passega C-M diagram of sediments below 140 cm in LW profile. C: Grain size value of the first percentile (the coarsest) on cumulative curve; M: Median grain size; T: Pelagic suspension; OP: Rolling & Bottom suspension; PQ: Rolling & Graded suspension; QR: Graded suspension; RS: Uniform suspension.

Fig. 6 Grain size distributions of typical samples in the LW profile.

Fig. 7 Comparison of TOC in sediments of LW profile with stalagmite $\delta^{18}\text{O}$ record (Wang et al., 2001).

Fig. 8 Time-series of TOC, $\delta^{13}\text{C}_{\text{org}}$, fern spore and tree pollen percentages in the LW profile.

Fig. 9 Comparison of TOC contents in LW profile with stalagmite $\delta^{18}\text{O}$ of Hulu Cave (Wang et al., 2001), ^{10}Be record in loess in northern China (Zhou et al., 2012) and stalagmite $\delta^{18}\text{O}$ record in Yemen (Shakun et al., 2007). The grey dashed lines show the monsoon strength at around 16,000 a BP.

Fig. 10 Comparison of TOC in sediments of LW profile and stalagmite $\delta^{18}\text{O}$ of Hulu Cave with sedimentary $^{231}\text{Pa}/^{230}\text{Th}$ in North Atlantic (McManus et al., 2004), ΔSST of the Eastern (Dubois et al., 2011) and the Western Tropical Pacific (Mohtadi et al., 2010). The gray shaped area shows the time span of the H1.

Table 1 AMS ^{14}C ages of LW profile

Laboratory code	Sample number	Dated material	Depth (cm)	Conventional age (a BP)	Calibrated age (a BP)
10442	LW-4	Bulk organic carbon	4	1,120±25	1,030±40
10443	LW-102	Bulk organic carbon	102	6,530±30	7,450±30
19041	LW-142	Bulk organic carbon	142	8230±30	9200±60
18731	LW-173	charcoal	173	10040±30	11550±120
10444	LW-173	Bulk organic carbon	173	9,840±40	11,250±40
10445	LW-215	Bulk organic carbon	215	4,500±30	5,170±90*
10446	LW-249	Bulk organic carbon	249	12,760±40	15,070±80
10447	LW-282	Bulk organic carbon	282	7,100±30	7,930±40*
10448	LW-371	Bulk organic carbon	371	16,810±45	19,970±90
10449	LW-435	Bulk organic carbon	435	22,610±70	26,850±60

454 *Samples close to covering soil of the profile are likely polluted by “new carbon” and
 455 thus deleted.

Figure 1

Figure 1_black-white

Figure 2

Figure 2_black-white

Figure 3

Figure 4

Figure 5

Figure 6

Figure 6_black-white

Figure 7

Figure 7_black-white

Figure 8

Figure 9

Figure 9_black-white

Figure 10

Figure 10_black-white