Extraordinary flood events and the response to monsoonal climatic change during the last 3000 years along the middle Yangtze River valley, China

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ABSTRACT

Well-dated long-term flood data series is very important in studying of hydrological response to past climatic change along the mainstream of the Yangtze River. Five palaeoflood events recorded by slackwater deposits were identified by the sedimentological criteria and analytical results. A detailed chronological framework of palaeoflood events during the late Holocene was established by optically stimulated luminescence dating in combination with anthropogenic remains, and with pedo-stratigraphic correlations with well dated Holocene loess-soil profiles within the middle Yangtze River basin. The five palaeoflood events were dated to 2.43 ± 0.18 ka, 1.82 ± 0.17 ka, 0.94 ± 0.07 ka, 0.72 ± 0.06 ka and 0.59 ± 0.05 ka, respectively. These results are consistent with geological records at other palaeoflood sites and historical flood records along the middle Yangtze River valley. A regional synthesis of the palaeoflood chronology over the last 3000 years was compiled along the middle Yangtze River. Extraordinary flood events generally seem to be correlated with late Holocene climatic variability (i.e. Roman Warm Period, Dark Ages Cold Period, Medieval Climate Anomaly, and Little Ice Age). The high-resolution climatic proxies from stalagmites of the Dongge cave and Heshang cave, ice-cores of the GRIP, the total solar irradiance variations and sunspot number, and *ENSO* activities suggest that these hydroclimatic events are possibly related to the weaker Asia summer monsoon and the cooling climatic events, as well as the stronger *ENSO* activities during the late Holocene in the Yangtze River valley. These results provide insights into the response of hydroclimatic system to global change in the large Rivers of Asia.

Keywords: palaeoflood events; OSL dating; late Holocene; climatic variability; middle Yangtze River

1. Introduction

Palaeoflood events provide a framework for more reliable interpretation of short-term instrument records, which are important archives of information in understanding extreme hydrologic event response to global climatic change (Knox and Kundzewicz, 1997; Knox, 2000). Sedimentological records of hydrological extremes can be complemented with historical documentary information and

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botanical records (Benito et al., 2015ab). A long-term perspective on hydrographical change was recorded by slackwater deposits (SWDs). These SWDs are typically fine-grained sand and silt that accumulate rapidly from suspension during major floods in the slackwater (back water) areas where current velocity is reduced (Baker et al., 1983; Ely and Baker, 1985; Benito et al., 2003). A series of sedimentological criteria can be used for identifying palaeoflood SWD in the field (Baker and Kochel, 1988; Benito and Thorndycraft, 2005). In our previous studies, extraordinary flood SWDs were interbedded in the Holocene loess-soil profiles at many sites along the middle reaches of the Yellow River and its tributary Rivers (Huang et al., 2007, 2010, 2011, 2012ab).

Palaeoflood investigations had been carried out on the Three Gorges reach and the Jiangnan Basin along the Yangtze River since the 1990s (Zhan and Xie, 1997, 2001; Yang and Xie, 1997; Zhu et al., 1997, 2005, 2008; Yang and Wang, 1997; Ge et al., 2004; Ge Z.S, 2009). The chronology of these palaeoflood events were revealed by radiocarbon that dating on wood, charcoal, buried soil and fine-organic detritus preserved in flood sediments in the lower Yangtze River (e.g., Yu et al., 2003). However, obtaining age control has been difficult due to incorporation of old organic matter into the sedimentation cycle in many fluvial sediments, which limits the application of ^{14}C dating (Rittenour, 2008; Fuchs et al., 2010). Over the last decade, optically stimulated luminescence (OSL) technology has been directly used for dating Holocene fluvial deposits, including floodplain deposits, palaeochannels, and palaeoflood sediments (e.g., Olley et al., 1998; Stokes et al., 2001; Zhang et al., 2003; Rodnight et al., 2006; Williams et al., 2010; Medialdea et al., 2014). Geochronological control for extraordinary floods and their palaeoclimatic implications have been established by OSL dating in the world's great river systems (Kale et al., 2000; Macklin et al., 2003; Thomas et al., 2007; Benito et al., 2011; Wasson et al., 2013; Grodek et al., 2013; Greenbaumet al., 2006; 2014ab).

Palaeoflood studies have suggested that extreme floods are sensitive indicators of climate-related changes in large river systems worldwide (e.g., Ely et al., 1993; Ely, 1997; Thorndycraft and Benito, 2006; Baker, 2008; Jones et al., 2010). East Asia, with a quarter of the world's population, suffers from recurrent floods and droughts and is highly vulnerable to extreme climate change (Lim and Fujiki, 2011). The Yangtze River, the longest river in Asia, is an area sensitive to monsoon climatic change. Chronology of extraordinary flood events in the Yangtze River is too short to adequately understand the extreme hydrological event response to monsoon climatic change. To our knowledge, palaeoclimatic analysis on extraordinary flood events during late Holocene are still lacking in the mainstream of the Yangtze River valley. This paper presents our new palaeoflood investigations in the

middle Yangtze River valley. These identified palaeoflood events will provide insight into the regional response of the hydro-climatic system to global change in the largest River of Asia.

2. Geographical setting

The Yangtze River valley is formed in response to regional extension throughout eastern China, synchronous with the start of strike-slip tectonism and surface uplift in eastern Tibet (Wang et al., 2013; Zheng et al., 2013). The study area is located at the margin of the loess distribution in China (Liu, 1988; Fig. 1A). Aeolian sediments (e.g., Wushan loess) have previously been identified in the middle reaches of the Yangtze River valley (e.g., Zhu et al., 1997; Jia et al., 2012; Hu et al., 2013; Wu et al., 2014). Typical loess is distributed in the north of the study area (Loess Plateau). The southern the study area is red clay, which is mainly located in the lower and middle reaches of the Yangtze River in subtropical China (e.g., Hu et al., 2010). The Xiashu loess is distributed in the eastern study area, which is mainly located in the lower reaches of the Yangtze River (e.g., Hao et al., 2010). The Ganzi loess is distributed in the western study area, which is mainly located at the west of Sichuan Province (e.g., Yang et al., 2010).

The mainstream of the Yangtze River is about 6300 km in length with a drainage area of 1.80 million km². It is the longest river in Asia and the third one in the world in terms of the length and discharge (Chen et al., 2001ab; Yu et al., 2009; Fig. 1B). The Yangtze River crosses the three major morphological steps from the Tibet plateau to the eastern coast (e.g. Yu et al., 2009; Fig. 1B). The upper reaches of the Yangtze River primarily run across mountainous terrain from the source to Yichang. The middle reaches of the Yangtze River lie between Yichang and Hukou. Three Gorges Dam (TGD) and Gezhouba (GZB) Dam are located at the boundary between middle and lower reaches of the Yangtze River (Chen et al., 2001ab; Fig. 1B). In the Three Gorges reach, the estimation of the largest palaeoflood, historical flood and gauging flood at the Yichang hydrological station were 114000, 105000 and 70800 m³ s⁻¹, respectively (Zhan and Xie, 2001; Ge, 2009).

Palaeoflood investigations were carried out along the upper and middle reaches of the Yangtze River during 2010–2013. The study area crosses two basins from Sichuan Basin to Jiangnan Basin, which is situated in the humid zone with a subtropical monsoonal environment in China (Fig. 1B and C). The mean annual temperature is 16.9°C and the mean annual precipitation 1200–1600 mm, which is brought about by the southeastern and southwestern maritime monsoons. The regional precipitation and hydrology are characterized by high seasonal variability with 70%–80% of rainfall occurring from June to October. Wujiang River, Jialingjiang River and the Three Gorges reach of the Yangtze River

are one of important flood sources in the middle reach of the Yangtze River (Fig. 1C).

3. Sampling site and methods

3.1. Sampling site

Detailed geomorphological, sedimentological and stratigraphic observations were carried out at the WBC site. The study site lies in the transition zone between Wushan Mountains and Jiangnan Basin. The five river terraces, respectively, are 10–14 m, 20–27 m, 35–42 m, 55–75 m and 95–105 m above the present river level in the Yichang reach of the Yangtze River. The first terrace is an alluvial terrace and the other terraces are bedrock terraces (Gu et al., 1983). The Holocene loess-soil stratigraphy was surveyed in detail adjacent the Weibicun (WBC) village of Yidu city, Hubei province (Fig. 1D). Two large dams (TGD and GZB Dam) are located upstream of the study site (Fig. 1C). The WBC site is located on the first alluvial terrace. The front of the WBC profile is about 10 m above the normal water level (~41 m) of the Yangtze River (Fig. 2). The Chadian station is located upstream ~3 km of the study site. The warning water level and safety water level at the Chadian station are 51.5 m and 53.3 m, respectively. The largest gauged flood, with a discharge of $53000 \text{ m}^3 \text{ s}^{-1}$, was recorded on August 17th 1998. The most recent large flood occurred on July 20th, 2010 with a discharge of $41400 \text{ m}^3 \text{ s}^{-1}$ at Chadian station (Fig. 2). The elevations of top surfaces of palaeoflood SWDs only provide a minimum estimation of palaeoflood stages (Jarrett and England, 2002; Fig. 2). Because the studied reach is an alluvial reach, the estimation of palaeoflood discharge is not recommended (Baker and Kochel, 1988). These SWDs were deposited by palaeofloods with a minimum magnitude close to the largest gauged floods of 1998 and 2010 (53000 and $41400 \text{ m}^3 \text{ s}^{-1}$) and seen between 48.65 and 53.29 m asl in the riverbank. The riverbanks between 48 and 51 m asl represent a threshold level for registration of extraordinary overbank floods. In the studied WBC profile at an elevation of 51 m asl on the right riverbank, several SWD bedsets were found within the Holocene loess-soil stratigraphy (Fig. 2). Thus, these SWDs were defined as the sedimentary record of extraordinary overbank flood events that occurred during the Holocene.

3.2. Methods

Pedo-stratigraphic subdivisions were observed during the fieldwork by close examination of the color, texture and structure in the WBC profile (Munsell Color Company, 2000; Schoeneberger et al., 2002). The Holocene loess-soil profiles with five palaeoflood SWDs interbeds were identified by sedimentological criteria used extensively in the world of palaeohydrology (Baker et al., 1983; Baker and Kochel, 1988; Benito et al., 2003; Benito and Thorndycraft, 2005; Huang et al., 2007, 2010, 2011,

2012ab). Detailed pedo-stratigraphic subdivisions and descriptions of the sediments in the WBC profile were shown in Table 1. Sediment samples (including 5 palaeoflood SWDs and 6 loess-soil samples) were systemically taken in the WBC profile. In addition, the 2010 and 2012 flood SWDs were collected at the WBC site for comparison with palaeoflood SWD in the same reaches (Fig. 1D). Ten OSL samples (including 4 palaeoflood SWDs, 5 loess-soil samples and 1 pottery sample) were taken at the WBC site.

OSL laboratory preparation included treatment with HCl (10%) and H₂O₂ (30%) to remove carbonates and organics matter, and dry sieving to isolate grains of 63–90 µm in size. The 63–90 µm grain fractions was etched with 40% HF for about 40 minutes to remove all feldspar and the surfaces of the grains, and then treated with 10% HCl to remove acid-soluble fluoride precipitates (Zhou et al., 2009). The purity of quartz grains was examined by detecting the infrared-stimulated luminescence (IRSL) signal and the shape of the 110°C TL peak (Mejdahly and Christiansen, 1994). Small-aliquots composed of ~200 grains, were made by spraying silicone oil on to 10-mm-diameter stainless steel disks through masks with hole diameters of 2 mm. A monolayer of quartz grains was then adhered to the arrayed portion of each disk surface. OSL measurements were made using a Risø TL/OSL-DA-20 reader. Blue light (470 ± 30 nm, ~50 mW cm⁻²) LED stimulation was used for the quartz grain OSL measurements. The OSL signal was detected with an EMI9235 QB15 photomultiplier tube with a 7-mm-thick Hoya U-340 glass filter. The reader was also equipped with a ⁹⁰Sr/⁹⁰Y beta source delivering 0.081 Gy s⁻¹ and an infrared laser diode (830 nm, ~145 mW cm⁻²). The equivalent doses (D_e) were measured using the standard single-aliquot regenerative-dose (SAR) procedure (Murray and Wintle, 2000). At least 24 aliquots of coarse grain quartz were measured from each sample. After subtracting the background estimated from the last 0.32 s integral of a 40 s stimulation, the integral of the 0–0.64 s section of the shine-down curves was used for D_e determination. The U, Th and K contents of the samples were determined by neutron activation analysis in the China Institute of Atomic Energy, Beijing. Water content (mass of moisture/dry mass; Aitken, 1998) was estimated by weighing the samples before and after drying (the samples were dried at 105°C for 24 h). The calculation was performed in Age.exe (Grün, 2003). Grain-size distribution was analyzed using the Malvern Mastersizer-S Particle Sizer with (NaPO₃)₆ as a dispersing agent after pretreatment with 10% HCl to remove secondary carbonates and 10% H₂O₂ to remove the organic matter (Huang et al., 2013).

4. Results

4.1. Sedimentary record of palaeoflood events

On the river banks along the mainstream in the middle reaches of the Yangtze River, the sandy silt SWD beds were found interbedded in the Holocene loess-soil profile at the WBC site. Palaeoflood SWD beds recording individual flood events were identified by close examination of the abrupt change in color, texture and structure in the WBC profile (Fig. 2; Table 1). The shape of particle-size distributions (PSD) is an available proxy for distinguishing palaeoflood SWD and other kinds of sediment (Huang et al., 2012; Guo et al., 2015). The PSD curves of SWD1 and SWD2 peak at 25–40 and 40–60 μm , respectively. The PSD curves of SWD3, SWD4 and SWD5 peak at 25–30, 35–70, and 35–40 μm , respectively (Fig. 3A). These palaeoflood SWDs are well-sorted, sandy silt. In contrast, loess deposits are clayey silt. The PSD curves of loess (L_0), palaeosol (S_0) and modern soil (MS2) peak at 20–30, 8–12, 15–25 μm , respectively (Fig. 3B). The intensity of weathering and pedogenesis suggest that clay and fine silt in the palaeosol S_0 are higher than those of loess L_0 in the WBC profile (Figs. 2 and 3). The character of aeolian deposits has been identified by particle size analysis of loess deposits in the middle Yangtze River (Li et al., 2010; Hu et al., 2013; Wu et al., 2014). Loess deposits at the WBC site show the features of a eolian deposits with fine silt and clay, which is different from these flood deposits (Figs. 2 and 3).

The particle-size distribution is a major proxy in studying sediment characteristics, the source of sediments, transporting force and sedimentary environment (Huang et al., 2010; Guo et al., 2014). The curves of the 2010 and 2012 flood SWDs peak at 7–12 μm at the WBC site, whereas the curves of the palaeoflood SWDs peak at 25–60 μm (Fig. 3A). Obviously, the particle-size distribution of modern flood SWDs are finer than that of the palaeoflood SWDs at the WBC site. Modern flood SWDs are fine silt, whereas palaeoflood SWDs are sandy silt. The reason is that the Three Gorges Dam and Gezhouba Dam locate on the upstream of sampling site, and the majority of bedload sediments entering the Three Gorges Reservoir was trapped (e.g. Wang et al., 2009; Li et al., 2011; Fang et al., 2012; Yuan et al., 2012; Fig. 1C). The PSD curves are characterized by a sharp peak in modern flood SWDs and palaeoflood SWDs.

Furthermore, particle-size distributions and the statistical indices of palaeoflood SWDs in the WBC profile were also presented in Table 2. Palaeoflood SWDs are dominated by sand (> 63 μm ; 34.7–63.3 %) and coarse silt (16–63 μm ; 22.9–40.0 %). Modern flood SWDs are dominated by fine silt (2–16 μm ; 32.0–34.6%) and coarse silt (44.2–49.0%). Loess deposits (including MS, L_0 and S_0) are dominated by fine silt (21.2–34.9%) and coarse silt (39.3–48.3%). The sand content in the

palaeoflood SWDs is higher than those of loess deposits. On the contrary, clay content ($< 2 \mu\text{m}$; 5.5–8.8%) in the palaeoflood SWDs is lower than those of loess deposits (9.0–15.6%). The values of medium size and mean size in the palaeoflood SWDs are obviously larger than those of loess deposits in the WBC profile. Thus, palaeoflood SWDs are well defined by an integration of field observations and analytical results (Fig. 3; Table 2).

4.2. Stratigraphy and chronology of palaeoflood events

4.2.1. Luminescence dating

Typical natural OSL decay curves and dose–response curves for coarse quartz fractions from samples WBC–B2 and WBC–B5 are shown in Fig. 4. It is clear that the OSL signal decreases very quickly during the first second of stimulation (Fig. 4A and B), suggesting the dominance by the fast OSL component. The OSL signals produced by zero doses are close to zero, suggesting that thermal transfers during measurement are negligible. These luminescence properties indicate that all the samples are suitable for D_e determination using the SAR protocol.

A preheat plateau experiment was carried out on samples WBC–B2 and WBC–B5, with preheat temperatures increasing from 180 to 300°C at 20°C intervals. At least three aliquots were measured using the SAR protocol at each temperature, with a heating rate of 5°C s⁻¹. A pre-heating plateau was observed from 220°C to 300°C in the sample WBC–B2 (Fig. 4C), and from 180°C to 240°C in the sample WBC–B5 (Fig. 4D). Based on the observations, the D_e shows a greater magnitude of the thermal transfer effect at temperature above 240°C (Fig. 4E). This is very likely to be related to thermal transfer occurring in the lattice. However, the effect of thermal transfer on the dose for the sample WBC–B2 and WBC–B5 at 220°C was 0.11 Gy and 0.15 Gy respectively. When compared with laboratory dose, the thermal transfer effect of the two samples at 220°C is both $< 5\%$ (Fig. 4E). In addition, the dose recoveries tests show that the ratios of measured dose to given dose are both in the range of 0.90–1.10 (Fig. 4F). The recycling ratios are consistent with 1.0 within 1 sigma from 180°C to 240°C. As a result, a preheat temperature of 220°C and a cut-heat of 180°C were used for routine D_e determination of the palaeoflood samples and aeolian samples.

The histograms distribution, cumulative frequency distribution and the radial plots for aeolian samples (WBC–B5) and palaeoflood SWD samples (WBC–B2, B3 and B6) were shown in Fig. 5. The over-dispersion (OD) is defined as the scatter that is useful for comparing the spread in the distribution of D_e between samples (Galbraith et al., 1999). The D_e distributions obtained for these samples vary markedly in type and form. The variety and complexity of these D_e distributions indicates that the

samples at the WBC site have been subjected to a diverse array of bleaching conditions and post-depositional processes. (e.g. Olley et al., 1999; Lepper et al., 2000; Stokes et al., 2001; Fuchs and Wangner, 2003; Zhang et al., 2003; Rittenour, 2008; Arnold et al., 2009; Fuchs et al., 2010; Summa-Nelson and Rittenour, 2012). In order to determine the true burial doses of the samples, the D_e values used for age calculation were analyzed using the central age model (CAM) and minimum age model (MAM) (Galbraith et al., 1999). A parameter of σ_b , which represents the scatter that was not explained by heterogeneous bleaching, was required in the model. The OD calculated using the D_e data from the dose recovery test for sample WBC-B2 is $11 \pm 4\%$. However, because of instrument errors and other intrinsic sources of variability, well-bleached natural samples should show greater variability than do the results from the dose-recovery experiments (Thomsen et al. 2007; Galbraith & Roberts 2012; Guo et al., 2016; Zhou et al., 2016). Thus, we chose a dispersion threshold (σ_b) of 15% for all samples, which was greater than the measured values. The central and minimum doses obtained for these samples are displayed in Fig. 5 and listed in Table 3. The coarse-grain fractions of all samples at the WBC site yielded ‘central’ OSL ages of from 6.00 to 0.6 ka, and ‘minimum’ OSL ages range from 5.73 to 0.38 ka. Therefore, the application of the CAM and MAM approaches to all samples allows more accurate estimates of the equivalent dose and age.

4.2.2 Chronology of palaeoflood events

Results for OSL ages (based on the MAM ages) in the WBC profile are in correct stratigraphic order (within uncertainties) and a linear fit of ten data points provides a total accumulation rate of ~ 1.05 mm per year (Fig. 6). Five palaeoflood SWDs interbedded in Holocene loess-soil stratigraphy agree with anticipated rapid deposition at the depth of 250–0 cm and 450–250 cm, respectively (Fig. 6). The pedo-stratigraphic subdivisions at the WBC site in the Yangtze River were well correlated to those identified at the Qianfangcun (QFC) site (Pang et al., 2011; Gu et al., 2012; Huang et al., 2013) in the Hanjiang River (Fig. 7). During the mid-Holocene Climatic Optimum (8500–3100 a B.P.), the palaeosol (S_0) developed on this terrace at the WBC site (Fig. 2). The loess L_0 of late Holocene age blanketed the palaeosol S_0 during the period 3000–1500 a B.P. in the WBC profile (Fig. 7). Two palaeoflood SWDs interbedded within the loess L_0 were found in the depth range of 450–250 cm. The first palaeoflood event recorded by SWD1 was dated to 2.43 ± 0.18 ka in the depth range of 410–400 cm. The bottom boundary of palaeoflood SWD2, in which the pottery shards and fragments of bricks and segmented tiles of the Han Dynasty (2156–1850 a B.P.) were retrieved from the depth range of 300–310 cm, was dated to 1.82 ± 0.17 ka. Three palaeoflood SWDs interbedded in the modern soil

were within the depths of 250–0 cm. The buried soil horizon MS4, MS3, MS2 and MS1 were dated to 1.26 ± 0.08 ka, 0.72 ± 0.09 ka, 0.62 ± 0.08 ka, and 0.38 ± 0.04 ka, respectively. The top part of palaeoflood SWD3, SWD2 and SWD1 was dated to 0.94 ± 0.07 ka, 0.72 ± 0.06 ka and 0.59 ± 0.05 ka, respectively (Fig. 8). Thus, the five palaeoflood events were dated to 2.43 ± 0.18 ka, 1.82 ± 0.17 ka, 0.94 ± 0.07 ka, 0.72 ± 0.06 ka and 0.59 ± 0.05 ka, respectively.

5. Discussion

Stratigraphic evidence of large Holocene flood is well preserved in the loess-soil sequences at the WBC site in the middle reach of the Yangtze River valley. A detailed chronological framework of late Holocene palaeoflood events at the WBC site was established by OSL dating and checked by anthropogenic remains and pedo-stratigraphic correlations with well dated Holocene loess-soil profile within the middle Yangtze River basin (Figs. 6 and 7). In the previous studies, palaeoflood investigations have been broadly carried out along the upper and middle reaches of the Yangtze River (e.g., Zhu et al., 1997, 2005, 2008; Xie et al., 2007; Fig. 8). Based on archaeological culture layers and stratigraphic correlation, late Holocene palaeoflood chronologies were established at the archaeological ruins (e.g., Zhongba site, Yuxi site and Zhongqiao site) in the Three Gorges reach and the Jiangnan plain of the Yangtze River during the Western Zhou Dynasty (BC 1046–771), Eastern Zhou Dynasty (BC 770–256), Tang Dynasty (AD 618–907), Song Dynasty (AD 960–1279) and Qing Dynasty (AD 1644–1911) in Chinese dynasties (Zhu et al., 1997, 2005, 2008; Wu et al., 2015; Fig. 8A). Sedimentary record of palaeoflood events were also identified by sand content at the Jiangling site within the Jiangnan basin (Xie et al., 2007; Fig. 8A). Late Holocene palaeoflood chronology (23 data) was established by ^{14}C dating in the Three Gorges reach of the Yangtze River (Zhan and Xie, 2001; Fig. 8B). A regional synthesis of the palaeoflood chronology (in total 28 data points) including 23 ^{14}C dates (data from Zhan and Xie, 2001) and 5 OSL dates from this study over the past 3000 years was compiled in Fig. 8C. Thus, the new optical chronology at the WBC site in the middle reach of the Yangtze River valley not only extends stratigraphic evidence of palaeoflood events, also shows good correlation with sedimentary record of palaeoflood at the other sites along the Yangtze River (Fig. 8).

Five major clusters of palaeoflood events (red numbers in parentheses) in the middle Yangtze River were constructed by the regional palaeoflood databases (in total 28), and occurred at 2600–2200, 1800, 1200, 1000–800, 600 a BP. (Figs. 8C). These flood phases generally correspond to the periods of the Eastern Zhou Dynasty (BC 770–256), Eastern Han Dynasty (AD 25–220), Tang Dynasty (AD 618–907), Song Dynasty to Yuan Dynasty (AD 960–1368) and Ming Dynasty (AD 1368–1644) in

Chinese history (Figs. 8 and 9). Evidence from historical documents in the Hubei Province within the middle Yangtze River basin indicated that periods of large historical floods were consistent with the five major clusters of palaeoflood events (China Meteorological Disasters Books, 2007; Fig. 9). A careful literature survey illustrated that lake expansion periods (BC 500–AD 0, AD 650–950 and AD 1250–1650) were due to climatic fluctuations and correspond to frequent flood reports (Fang, 1993; Fig. 9). In the subtropical monsoonal regions, abnormal climate variability is typified by extraordinary floods resulting from rainstorms and severe droughts resulting from shortage of rainfall (Huang et al., 2010; 2013). Our previous palaeoflood studies highlight that both floods and droughts were parts of climate variation (Huang et al., 2011, 2012 a b; Guo et al., 2015). The records of climatic anomalies (i.e., floods, droughts) in the ancient Chinese documents commonly displayed abrupt climatic changes during the 2000 years of historical documents (Ge et al., 2012, 2013; Zheng et al., 2014). Palaeoflood events occurred during the periods, in which severe droughts also occurred in the study area (China Meteorological Disasters Books; Fig. 9). For example, the drought periods coincide with fire events (i.e. 2600–2400, 1400–1200, 1000–800 and 600–400 a B.P.) during late Holocene were reconstructed in the southeastern China (Huang et al., 2014). Thus, palaeoflood events cluster (red numbers in parentheses) during the late Holocene in the middle Yangtze River valley and seem to be correlated with late Holocene climatic variability, such as Roman Warm Period (RWP), Dark Ages Cold Period (DACP), Medieval Climate Anomaly (MCA), and Little Ice Age (LIA) (Figs. 8 and 9).

The five palaeoflood event clusters (2600–2200, 1800, 1200, 1000–800, 600 a B.P.) over the last 3000 years in the middle Yangtze River were considered as a part of global hydroclimatic variability during the late Holocene. In turn, these palaeoflood event clusters were also synchronous with global hydroclimatic changes. Holocene palaeohydrology recorded by flooding activity occurred during 2800–2400, 1500–1300, 1200–1050, 900–800, and 700–200 a B.P.. Over the last 3000 years, these periods match with glaciers advancing to their maximum extent from Mont-Blanc Glacier and high lake levels in west-central Europe (Holzhauser et al., 2005; Debret et al., 2010; Giguët-Covx et al., 2012). Rapid ice cap growth and landscape instability (i.e., 3000, 2600, 1400 and 700 a B.P.) were revealed by a 3000-year varved record of glacier activity and climate change from the proglacial lake Hvítárvatn, Iceland (Larsen et al., 2011). Four cooling events during late Holocene were distinguished from core MD99-2271 and MD99-2269 on the North Icelandic shelf, centered at about 2600, 1900, 1300 and 600 a B.P., respectively (Jiang et al., 2006, 2007; Sha et al., 2014). Thus, the five palaeoflood clusters provide important advances in interpreting the local and regional response of

hydro-climatic system to global change from the largest River in Asia.

Palaeoflood clusters are closely related to the lower $\delta^{18}\text{O}$ values from stalagmites, which reflect general weakening of monsoon activities (red numbers in parentheses). High resolution climatic proxies and precise dating from stalagmites can provide information on shifts in monsoon precipitation (Wang et al., 2005; Hu et al., 2008). However, some differences between the Dongge and Heshang Cave imply that the Heshang Cave probably records regional precipitation changes (Fig. 10A and B). Palaeoflood event clusters in the middle Yangtze River valley also show good agreement in terms of the low values of $\delta^{18}\text{O}$ from ice-cores of the GRIP (Dansgaard et al., 1993; Fig. 10C). A number of studies have argued that solar activity is a fundamental forcing producing the Eastern Asia monsoon climatic variability (e.g., Wang et al., 2005; Sagawa et al., 2014). Extraordinary flood event clusters during the late Holocene in the middle Yangtze River agree well with a weaker total solar irradiance and a lower sunspot number (Steinhilber et al., 2009; Fig. 10D and E). Furthermore, the palaeoflood event clusters in the Yangtze River valley appear to be associated with stronger El Niño/Southern Oscillation (*ENSO*) activities (Moy et al., 2002; Fig. 10F). In the previous studies, variability and possible teleconnections between floods and *ENSO* events had also been investigated in the Yangtze River system (e.g., Shi et al., 2004; Jiang et al., 2004, 2006; Zhang et al., 2007, 2008). Thus, these climatic proxies broadly show that sharp hydrological changes in the Yangtze River valley are possibly consistent with the weaker Asia summer monsoon and the cooling climatic events during the late Holocene, as well as the stronger *ENSO* activities.

How do we assess the link between extreme flood events and abnormal monsoon circulation (weak monsoon activity) in the Yangtze River basin? Modern hydrometeorologic records may provide an important opportunity for understanding the nature of extreme hydrologic events in the context of changing monsoon circulations (e.g., Ely, 1997). East Asia Summer Monsoon (EASM) and Indian Summer Monsoon (ISM) controlling atmospheric circulation during the summer largely occur in location of the “July Jet Stream” and Siberian-Mongolian high controlling atmospheric circulation during the winter mainly occurs in location of the “January Jet Stream” (Peng et al., 2005; Fig. 11A). However, when the weak summer monsoon occurs in the Yangtze River catchment, the warm-humid air masses brought by EASM and ISM meets the northern cold airflow brought by northwestern continental monsoon. The quasi-stationary front hovers above the upper and middle Yangtze River catchment, which would be likely to bring long-duration rainstorms and cause large-magnitude floods in this region. For example, during the most recently occurring large flood event in the Yangtze River

basin on July 20–23, 2012, the main sources of precipitation in the upper Yangtze River were brought by the abnormal monsoon circulation from the Indian Ocean (Yin et al., 2014; Fig. 11B). The magnitude of the 2012 flood is slight lower than reconstruction palaeoflood discharges during the Late Holocene in the Three Gorges reach of the Yangtze River. Peak discharges of the 2012 flood at the Zhutuo station and Cuntan station were $56500 \text{ m}^3 \text{ s}^{-1}$ and $67300 \text{ m}^3 \text{ s}^{-1}$, respectively (Fig. 11B). The highest flood peak ($71200 \text{ m}^3 \text{ s}^{-1}$) was recorded in the Three Gorges Reservoir inflow record (Yin et al., 2014). Peak discharges of palaeoflood events during the late Holocene in the Three Gorges reach were estimated to be $87200\text{--}102000 \text{ m}^3 \text{ s}^{-1}$ (Zhan and Xie, 2001; Ge et al., 2004; Ge, 2009). This gives us an opportunity to compare hydroclimatic variability between modern flood events caused by abnormal ISM circulation in the Yangtze River basin with the palaeoflood events (Fig. 11). These results may be helpful in understanding extreme flood and rainstorms events caused by the weak summer monsoon variability in the upper and middle Yangtze River.

6. Conclusions

Five palaeoflood SWDs interbedded Holocene loess-soil stratigraphy at the WBC site were identified by the sedimentary criteria and analytical results in the middle reaches of the Yangtze River. The chronology of extraordinary flood events over 3000 years was established by OSL dating and checked by anthropogenic remains and pedo-stratigraphic correlations. The five palaeoflood events were dated to $2.43\pm0.18 \text{ ka}$, $1.82\pm0.17 \text{ ka}$, $0.94\pm0.07 \text{ ka}$, $0.72\pm0.06 \text{ ka}$ and $0.62\pm0.05 \text{ ka}$, respectively. The OSL ages at the WBC site provide the first detailed chronology of extraordinary flood events during late Holocene in the mainstream of Yangtze River. Furthermore, a regional synthesis of the palaeoflood chronology over the past 3000 years was compiled in the middle reaches of the Yangtze River. The five major clusters of palaeoflood clutters occurred at 2600–2200, 1800, 1200, 1000–800, 600 a BP, which are in good agreement with the periods of Eastern Zhou Dynasty (BC 770–256), Eastern Han Dynasty (AD 25–220), Tang Dynasty (AD 618–907), Song Dynasty to Yuan Dynasty (AD 960–1368) and Ming Dynasty (AD 1368–1644) in Chinese history.

The reconstructed palaeoflood events clusters seem to be generally correlated with late Holocene climatic variability (i.e. Roman Warm Period, Dark Ages Cold Period, Medieval Climate Anomaly, and Little Ice Age). The high-resolution climatic proxies from stalagmites of the Dongge cave and Heshang cave, and ice-cores of the GRIP, total solar irradiance variations and sunspot number generally show that sharp hydrological changes in the Yangtze River are possibly consistent with the weaker Asia summer monsoon and the cooling climatic events during the late Holocene. The increase

of hydroclimatic events within the Yangtze River basin appear to be associated with stronger *ENSO* activities. The integration palaeoflood events over the last 3000 years provide a better interpretation for the regional response of hydroclimatic system to global change in the largest River of Asia.

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Figure captions:

Fig. 1. (A) Map showing the distribution of loess deposits (modified from Liu, 1988) and location of study site. (B) Map showing Yangtze River drainage basin and location of major hydrological gauging stations. (C) Dem map showing topographic features of the study site (■) and other mentioned palaeoflood sites (○) along the middle reaches of the Yangtze River valley. Box outlines the study area. (D) Detailed geomorphologic map of the Weibicun (WBC) site and the location of sampling site of modern flood deposits and Chadian station at in the middle reach of the Yangtze River valley.

Fig. 2. (A) Photo showing palaeoflood SWDs and Holocene loess-soil profile at the WBC site in the middle reach of the Yangtze River valley. (B) Photo showing the intensity of weathering and pedogenesis in palaeosol (S_0) in the WBC profile. (C) Simple diagrams showing that palaeoflood slackwater deposits (SWD) interbeds in Holocene loess-soil stratigraphy at the WBC site and

magnitude of modern large floods at the Chadian station.

Fig. 3. Comparison of the particle-size distribution frequency of palaeoflood SWDs with that of modern flood SWDs, loess (L_0), palaeosol (S_0) and modern soil (MS2) at the WBC site in the middle reach of the Yangtze River valley. (A) Particle-size distribution frequency curves show that palaeoflood SWDs are well-sorted sandy silt, whereas modern flood SWDs are well-sorted fine silt caused by the Three Gorges Dam and the Gezhouba Dam. (B) Loess deposits (including MS2, L_0 and S_0) are clayey silt at the WBC site, which are different from these flood deposits.

Fig. 4. Decay curves and growth curves, preheat temperature plateau test, thermal transfer test, the average Measured/Given ratio obtained at each preheat temperature for samples WBC-B2 and WBC-B5 at the WBC site in the middle Reach of the Yangtze River.

Fig. 5. The histograms distribution, cumulative frequency and the radial plots for samples WBC-B2, WBC-B3, WBC-B5 and WBC-B6 at the WBC site in the middle reaches of the Yangtze River valley.

Fig. 6. The relationship between OSL ages (based on MAM ages) and the depth in the WBC profile in the middle reach of the Yangtze River valley.

Fig. 7. Pedo-stratigraphic correlations between WBC site in the middle reach of the Yangtze River valley and the Holocene loess-soil profile at the Qianfangcun (QFC) site in the Hanjiang River valley (Huang et al., 2013).

Fig. 8. (A) OSL chronology of palaeoflood events at the WBC site has good correlation with the sedimentary record of palaeoflood events at the Jiangling site (Xie et al., 2007), Yuxi site and Zhongba site (Zhu et al., 1997, 2005, 2008) and the Three Gorges reach (Zhan and Xie, 2001) in the upper and middle reaches of the Yangtze River valley. (B) and (C) A regional synthesis of the palaeoflood chronology ($n=28$) including 23 ^{14}C dates (data from Zhan and Xie, 2001) and 5 OSL dates within the past 3000 years was compiled. Five palaeoflood events clutters (2600–2200, 1800, 1200, 1000–800, 600 a BP) generally seem to be correlated with late Holocene climatic variability, such as the Roman Warm Period (RWP), Dark Ages Cold Period (DACP), Medieval Climate Anomaly (MCA), and Little Ice Age (LIA). The red circles are OSL ages, the yellow parallelograms are ^{14}C ages, the blue rectangles are individual flood events, and the red numbers in parentheses are five palaeoflood events clutters.

Fig. 9. OSL chronology of palaeoflood events at the WBC site has good correlation with lake expansion periods (Fang, 1993) and large flood records in historical documents (China Meteorological Disasters Books, 2007). Historical floods and droughts records are parts of climatic variability in the

Hubei Province (China Meteorological Disasters Books, 2007). The red circles are OSL ages and the blue rectangles are lake expansion periods (Fang, 1993).

Fig. 10. The high-resolution climatic proxies confirm that the five extraordinary flood events (red numbers in parentheses) in the Yangtze River valley generally coincide with the lower values of the $\delta^{18}\text{O}$ values from stalagmites Dongge cave (Wang et al., 2005), Heshang cave (Hu et al., 2008) and ice-cores of the GRIP (smoothed 200yr, Dansgaard et al., 1993). Meanwhile, these palaeoflood events are consistently coincide with the lower values of the total solar irradiance variations (Steinhilber et al., 2009) and sunspot number (Solanki et al., 2004), but are related to the higher of the *ENSO* activities (Moy et al., 2002). Vertical gray bars denote the periods of palaeoflood event clutters (2600–2200, 1800, 1200, 1000–800, 600 a BP). The orange arrows denote the periods of palaeoflood events are possibly related to the weaker Asia summer monsoon and the cooling climatic events, as well as the stronger *ENSO* activities.

Fig. 11. A). Map showing the climatic system of China including East Asia Summer Monsoon (EASM), Indian Summer Monsoon (ISM) and Siberian-Mongolian high associated with the mean locations of jet stream (modified from Peng et al., 2005). Red line indicates rough boundaries between ISM and EASM (modified from Zhang et al., 2011). Yellow line indicates extraordinary flood events in the middle Yangtze River valley associated with the rough location of weak summer monsoon. B). The distribution of rainstorms in the Yangtze River basin on July 20–23, 2012 (modified from Yin et al., 2014) showing the main sources of the precipitation are brought by the abnormal monsoon circulation from ISM.

Table 1

Pedo-stratigraphic subdivision and description of the sediments in the Weibicun (WBC) profile in the middle reach of Yangtze River valley.

Depth (cm)	Pedo-stratigraphy	Pedo-stratigraphic descriptions
0–30	Modern soil (MS1)	Light brownish gray (7.5YR7/2), clayey silt, granular structure, loose porous, abundant plant roots and bio-pores, some earthworm burrows and excrement, well-rounded spherical pellets
30–70	Slackwater deposit (SWD5)	Dull yellow orange (10YR7/3), sandy silt, very loose, well-sorted
70–100	Modern soil (MS2)	Dull reddish brown (5YR5/3), clayey silt, granular structure, loess, some earthworm burrows and excrement, well-rounded spherical pellets
100–120	Slackwater deposit (SWD4)	Dull yellow orange (10YR7/3), sandy silt, very loose, well-sorted
120–150	Modern soil (MS3)	Dull reddish brown (5YR5/3), clayey silt, granular structure, loess, some earthworm burrows and excrement, well-rounded spherical pellets, few burnt earth
150–200	Slackwater deposit (SWD3)	Dull yellow orange (10YR7/3), sandy silt, very loose, well-sorted
200–250	Modern soil (MS4)	Dull reddish brown (5YR5/3), clayey silt, granular structure, loess, some earthworm burrows and excrement, well-rounded spherical pellets
250–300	Loess deposits (L_0)	Dull yellow orange (10YR6/4), silt, massive structure, few earthworm burrows and excrement, weak pedogenesis
300–310	Slackwater deposit (SWD2)	Dull yellow orange (10YR7/3), sandy silt, very loose, well-sorted
310–400	Loess deposits (L_0)	Dull yellow orange (10YR6/4), silt, massive structure, few earthworm burrows and excrement, weak pedogenesis, human remains of the Eastern Han dynasty (1900–1700 a BP), including rope-patterned gray pottery shards
400–410	Slackwater deposit (SWD1)	Dull yellow orange (10YR7/3), sandy silt, very loose, well-sorted
410–450	Loess deposits (L_0)	Dull yellow orange (10YR6/4), silt, massive structure, few earthworm burrows and excrement, weak pedogenesis,
450–750	Palaeosol (S_0)	Yellowish brown (2.5YR5/6), clayey silt, angular-blocky structure, relatively firm, density, abundant brown clay film in the crevice, strongly weathering, but no bio-pores

Table 2

Particle-size distribution of the palaeoflood SWDs in comparison with that of the modern (2010 and 2012) flood SWDs and loess deposits (MS2, L₀ and S₀) at the WBC site in the middle Yangtze River valley.

Sediments	< 2 μm (%)	2-16 μm (%)	16-63 μm (%)	> 63 μm (%)	Medium size (μm)	Mean size (μm)
Paleoflood SWD1	8.8	13.6	25.8	51.8	24.0	33.2
Paleoflood SWD2	8.5	13.3	35.8	42.4	20.4	24.6
Paleoflood SWD3	8.1	17.2	40.0	34.7	16.7	21.0
Paleoflood SWD4	5.5	8.3	22.9	63.3	34.4	44.2
Paleoflood SWD5	8.7	13.2	36.2	41.9	19.1	25.9
2012 flood SWD	10.2	32.0	49.0	8.7	8.9	9.3
2010 flood SWD	13.2	34.6	44.2	7.9	7.6	8.3
Modern soil MS2	9.0	21.2	48.3	21.5	12.5	14.8
Loess L ₀	11.2	21.4	41.1	26.3	12.5	15.1
Palaeosol S ₀	15.6	34.9	39.3	10.1	7.1	7.7

Table 3

Summary of sample depth, estimated water content, dose rates, estimates of dose over-dispersion (OD), and the equivalent dose values, and OSL ages for each sample at the Weibicun (WBC) site in the middle reach of the Yangtze River valley.

Sample ID	Depth (cm)	Material	U (ppm)	Th (ppm)	K (%)	Water (%)	N ^a	OD (%)	Dose rate (Ga/ka)	Central D _e (Gy)	Central age (ka)	Minimum D _e (Gy)	Minimum age (ka)
WBC-B1	25-30	MS1	2.84±0.11	11.3±0.32	1.61±0.05	18±5	32/38	41	2.76±0.18	1.66±0.12	0.60±0.06	1.05±0.07	0.38±0.04
WBC-B2	30-35	SWD5	3.3±0.12	11.3±0.32	1.71±0.06	18±5	32/38	27	2.92±0.19	2.02±0.10	0.69±0.06	1.71±0.07	0.59±0.05
WBC-A1	80-90	MS2	2.34±0.10	11.8±0.35	1.86±0.06	18±5	25/30	30	2.86±0.19	1.84±0.35	0.65±0.13	1.76±0.20	0.62±0.08
WBC-B3	100-105	SWD4	2.56±0.10	10.8±0.31	1.8±0.06	20±5	32/40	31	2.74±0.18	2.34±0.16	0.85±0.07	1.96±0.08	0.72±0.06
WBC-A2	120-130	MS3	2.70±0.10	10.2±0.33	1.70±0.06	20±5	33/45	33	2.65±0.17	2.13±0.32	0.83±0.13	1.90±0.20	0.72±0.09
WBC-B4	150-155	SWD3	2.41±0.10	10.7±0.31	1.72±0.06	23±5	30/36	35	2.56±0.16	3.22±0.20	1.26±0.11	2.40±0.11	0.94±0.07
WBC-B5	200-205	MS4	2.78±0.11	11.5±0.32	1.77±0.06	25±5	32/38	23	2.66±0.17	3.52±0.14	1.33±0.10	3.34±0.11	1.26±0.08
WBC-A3*	300-310	Pottery	2.43±0.10	12.5±0.38	1.93±0.06	25±5	18/20	11	2.78±0.15	4.77±0.31	1.82±0.17	-	-
WBC-B6	400-405	SWD1	2.85±0.11	9.68±0.29	1.65±0.06	25±5	30/38	34	2.44±0.16	6.41±0.21	2.63±0.24	5.93±0.20	2.43±0.18
WBC-A4	600-610	S0	3.35±0.11	13.8±0.40	1.93±0.06	25±5	35/45	32	2.95±0.20	17.70±1.30	6.00±0.60	16.90±1.10	5.73±0.53

^a accepted / measured.

* the sample WBC-A3 were calculated by common age model and other samples were calculated by central age model and/or minimum age model, respectively (Galbraith et al., 1999).

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