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Mid-late Holocene climatic changes recorded by loess deposits in the eastern margin of the Tibetan Plateau: Implication for human migrations

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ABSTRACT

The climatic fluctuations during the mid-late Holocene in northern Sichuan, southwest China, were inferred from loess deposits using multi-proxies of total organic carbon (TOC), carbonates, magnetic susceptibility, stable isotopic composition of organic carbon (δ^{13}Corg) and grain size distribution in combination with the dates determined by the accelerator mass spectrometry (AMS) ^{14}C and optically stimulated luminescence (OSL) methods. From ~6300 to 5500 a BP, the regional climate presented gradually increased warmth and humidity, subsequently characterized by reduced precipitation but with relatively high air temperature between ~5500 and 4800 a BP, demonstrating that the East Asian monsoon weakened. A warmer and wetter climate prevailed since ~4800 a BP and was interrupted by a sharp cold reversal at approximately 3300 a BP that was likely caused by solar irradiance forcing, which resulted in a global cold climatic change and glacier advance. Large-scale human migrations into the eastern margin of the Tibetan Plateau occurred in the cold and dry events that developed at ~6000, 2500 and 1200 a BP, indicating a critical stimulus effect of climatic change on human migration.

1. Introduction

Loess-paleosol sequences are key archives for paleoclimate and paleoenvironment reconstruction (Kukla, 1987; An et al., 1991; Liu et al., 1993; Ding et al., 1998; Guo et al., 2002). According to the loess evidence from the Loess Plateau, it is generally recognized that the cold and dry climate prevailed during the Early Holocene (before 9000 a BP), the warm and wet climate prevailed in the mid-Holocene (8000–3000 a BP) and the cold and dry climate has prevailed during the Late Holocene (since 3000 a BP) (Shi et al., 1993; Zhou et al., 1994; An et al., 2000; Huang et al., 2000, 2002; Xiao et al., 2002; Maher and Hu, 2006). Nevertheless, inconsistencies in the duration and the spatial pattern of climatic fluctuations, especially the epoch of the ‘warm/wet’ Holocene optimum and several abrupt cooling events, have been reported by various investigators (Shi et al., 1993; An et al., 2000; Xiao et al., 2002; Lu et al., 2011; Dong et al., 2012; Liu et al., 2013; Jiang et al., 2014; Xia et al., 2014; Wang et al., 2014; Sun and Feng, 2015; Yang et al., 2015).

The loess deposits are dotted extensively among the geographic positions of mountains and canyons in the eastern margin of the Tibetan Plateau, close to the Loess Plateau. They have been interpreted as a result of windblown dust accumulation derived from neighboring glaciofluvial outwash deposits based on the grain size distribution (Liu et al., 2007; Sheng, 2010a; Wen et al., 2014) and the surface microstructure of the quartz sands of those deposits (Fang et al., 1996; Wang et al., 2003; Sheng, 2010a; Wen et al., 2014). A colder and drier trend in climate in the eastern margin of the Tibetan Plateau since ~1.15 Ma BP has been inferred, based on the multi proxies of the loess deposits (e.g., magnetic susceptibility, carbonates, iron oxides, minor elements) (Jiang et al., 1997; Wang and Pan, 1997; Chen et al., 2002; Wang et al., 2005a, 2006). However, the Holocene climatic history inferred from the loess of this region was rarely reported. In the adjacent region of Qinghai Province, loess deposits recorded varied Holocene climate changes,

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e.g., during mid-late Holocene, wet conditions (5830–3940 a BP) was interrupted by a drought event (4900–4700 a BP), and the climatic change affected prehistoric cultural evolution (Dong et al., 2012); paleosol formation reflecting wet conditions developed at 5200-4000 a BP and 3900-700 a BP (Liu et al., 2012), whereas cold events occurred at 5300-4700 a BP, 3100-1500 a BP and 700-0 a BP (Liu et al., 2013). In terms of the peat records from the Hongyuan and Zoige regions in the eastern margin of the Tibetan Plateau, the climatic fluctuations existed with a centennial- to millennial- scale rhythm during the mid-late Holocene, and climatic reversals occurred around 6400–6200, 4400–4000, 3700–3500, 2800 and 1500 cal a BP (Zhou et al., 2002; Xu et al., 2002; Yu et al., 2006). A high-resolution stalagmite record from the Wanxiang Cave reflected Asian Monsoon variation over the past 1810 years that correlated with solar variability, alpine glacial retreat, and Chinese cultural changes (Zhang et al., 2008).

Human migration is closely linked to climate change (Zhang et al., 2011; Bayon et al., 2012; Kennett et al., 2012; Dong et al., 2013; Bohra-Mishraa et al., 2014; Chen et al., 2015a). Numerous substantial Neolithic sites/remains have been found in the eastern margin of the Tibetan Plateau, among which the earliest Neolithic relics (~6000 a BP) were undoubtedly later than those in the neighboring Gansu-Qinghai region (Jiang and Chen, 2001, 2003; Shi, 2006; Chen, 2007). With respect to the similarity in cultural character between these two regions, archaeologists commonly believe that the ancient inhabitants in the eastern margin of the Tibetan Plateau came from the Gansu-Qinghai region in northwestern China (Cheng and Wang, 2004; Sun and Deng, 2006; Shi, 2008, 2009). It was concluded that the early Neolithic sites before ~4600 a BP assembled in the upper reaches of the Min River region, and the late Neolithic sites, were mostly distributed in the Chengdu Plain (Fig. 1). The links of human migration during the mid- and Late- Holocene to climate change have not been interpreted due to the lack of satisfactory evidence from the integration between archaeological, historical and paleoclimatic investigations in these regions of northwest and southwest China.

Several archaeological relics dating back to the Neolithic Period and the Han Dynasty (~2000 a BP) were excavated from the loess deposits in the Jiuzhaigou Nature Reserve (JNR), which is a notable site for associating human migration with regional climate change in the eastern margin of the Tibetan Plateau. Here, we attempt to (1) reconstruct the mid-late Holocene climatic history in the JNR using the climate change proxies of total organic carbon (TOC), carbonates, ratio of free iron oxide (Fe_D) to total oxide (Fe_T), magnetic susceptibility, stable isotopic composition of organic carbon (δ^{13}Corg) and grain size distribution from loess deposits, and we identify evidence of the Holocene megathermal and several climatic reversals in combination with previous reports; (2) gain an understanding of the link between major human migrations and the climatic change over the past 6000 a BP in the eastern margin of the Tibetan Plateau.

2. Materials and methods

2.1. Sample collection in the JNR

The JNR is located in the northern Min Mountain region on the eastern margin of the Tibetan Plateau in northern Sichuan Province, China (Fig. 1). It is characterized by numerous canyons, and the terrain markedly descends northward from an elevation of 4789 m–1996 m above sea level (a.s.l.). The climate of the JNR is mainly controlled by the East Asian monsoon, which is characterized by the cold and dry northwesterly monsoon in winter and the warm and moist southeasterly summer monsoon. The mean annual precipitation in the region is between 550 mm and 780 mm and occurs primarily between April and August during the summer monsoon. The mean annual temperature is 7.3 °C, with an average of 16.8 °C in July and an average of −8.7 °C in January at an elevation of 2389 m a.s.l in the JNR.

The investigated loess-paleosol profile was excavated on a terrace in the JNR (Fig. 2) located at an elevation of ~2600 m a.s.l. (33°14′ N, 103°54′ E). The sampling profile was 5.7 m in depth, and two paleosols were identified at depths between 290 and 320 cm and between 485 and 525 cm below the surface. One hundred fourteen (114) bulk sediment samples were collected at intervals of 5 cm from the bottom to the surface of the profile. Seven (7) charcoal samples and 1 paleosol organic carbon sample were collected at a depth between 128 and 503 cm for radiocarbon dating, and an extra 6 samples were collected at a depth between 70 cm and 420 cm for optically stimulated luminescence (OSL) dating.

2.2. Laboratory methods

Total organic carbon (TOC) was measured by the Acid-Dichromate-$FeSO_4$ method, and $CaCO_3$ content was determined by treating the samples with HCl, and then calculated based on the evolved CO_2 (Soil Survey Staff, 1996). To determine the grain size distribution, the samples were first pretreated with H_2O_2 (30%) and HCl (10%) to remove organic matter and carbonates, respectively, and then measured with a Malvern Mastersizer 2000 Laser instrument. The free iron oxide (Fe_D) and total iron oxide (Fe_T) were extracted with dithionite-citrate-bicarbonate and HNO_3-HCl-H_2SO_4 solutions, respectively, and subsequently measured using phenanthroline spectrophotometry (Soil Survey Staff, 1996). The magnetic susceptibility was measured using a Bartington MS2 susceptibility meter at frequencies of 470 Hz and 4700 Hz after drying the samples at 38 °C for 48 h.

The loss of ignition and bulk composition of the major elements (SiO_2, Al_2O_3, Fe_2O_3, CaO, Na_2O, MgO, K_2O, and TiO_2) in 14 paleosol and 2 loess samples was determined using the Na_2CO_3 fusion method (Rettig et al., 1983) at the Institute of Mountain Hazards and Environment, Chinese Academy of Sciences.

The samples analyzed for organic carbon isotopic value ($\delta^{13}C$) were pretreated with 2 M HCl for 24 h to thoroughly digest carbonate carbon, then powdered and analyzed using a mass spectrometer (MAT-251) at the Institute of Earth Environment, Chinese Academy of Sciences. Values were expressed as per mil deviations relative to the Vienna Peedee Belemnite (VPDB) standard, with an analytical precision of ±0.2‰.

Radiocarbon dates were determined at the Xi'an AMS ^{14}C Center, Chinese Academy of Sciences using accelerator mass spectrometry and calibrated using a Calib601 program. The weighted mean ages were calculated using the 1σ probability distributions for each age (Data Repository, Table DR1). The OSL dating was performed at the School of Geographic and Oceanographic Science, Nanjing University (Data Repository, Table DR2).

3. Results

3.1. Chronology and sedimentation rate of the JNR profile

According to the linear regressions of our AMS ^{14}C and OSL dating ages (Fig. 3), the difference in evaluated ages between these two dating methods is less than 100 years in depths greater than 200 cm (i.e., the age exceeds ~ 1400 a BP), 100–150 years in depths between 200 and 100 cm (~700–1400 a BP) and 150–200 years in depths within 100–0 cm (~700–0 a BP). These are close to the analytical errors derived from the dating methods, showing that the AMS ^{14}C ages are in robust agreement with the OSL dating ages,

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