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Role of the mid-Holocene environmental transition in the decline of late Neolithic cultures in the deserts of NE China

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ABSTRACT

The mid-Holocene environmental transition was characterised by global cooling and the abrupt weakening of the Northern Hemisphere monsoon systems. It is generally considered the key driver of the collapse of several mid-Holocene agricultural societies, on a global scale. However, only a few previous studies have tried to verify the climatic origin of the collapse of these societies, using the compilation of spatiotemporal data at a large scale. Especially, the nature of mid-Holocene human-environment interactions in the climatically-sensitive margin of the East Asian summer monsoon front remains to be thoroughly understood. However, a systematic compilation of archaeological data at a regional scale can be used to verify the role the mid-Holocene environmental transition played in the collapse of late Neolithic cultures in China. Here, we present a regional compilation of Holocene records from sub-aerial sedimentary deposits, lake sediments, and archaeological sites in the deserts of NE China and the adjacent regions to explore human-environment interactions during the mid-Holocene. Comparison of the records of Holocene climate change with the evolution of archaeological sites reveals that the mid-Holocene environmental transition resulted in ecosystem degradation in the deserts of NE China, rendering these areas much less habitable. Faced with substantially increased environmental pressures, the late Neolithic inhabitants used several subsistence strategies to adapt to the environmental transition, including change in agricultural practices and ultimately migration. Overall, our results support the view that a widespread mid-Holocene drought destroyed the rain-fed agricultural and/or plant-based subsistence economies, ultimately contributing to the collapse of late Neolithic cultures in NE China.

1. Introduction

The global climate during the current Holocene interglacial is usually considered to be substantially warmer than that of the Younger Dryas event (Dansgaard et al., 1989). Moreover, it has provided a relatively stable environmental context for the growth and development of human society. However, when thresholds of the Earth system were crossed (Alley et al., 2003), abrupt and widespread climate changes with major impacts on social-ecological systems occurred repeatedly during the Holocene (Weiss et al., 1993; O'Brien et al., 1995; Alley et al., 1997; Cullen et al., 2000; deMenocal et al., 2000a; Thompson et al., 2002; Morrill et al., 2003; Staubwasser et al., 2003; Marchant and Hooghiemstra, 2004; Mayewski et al., 2004; Alley and Ágústsdóttir, 2005; Booth et al., 2005; Drysdale et al., 2006; Wanner et al., 2008, 2011; Moros et al., 2009; Zhao et al., 2009; Macdonald, 2011; de Boer et al., 2014; Sarkar et al., 2015; Shanahan et al., 2015; Solomina et al., 2015; Lillios et al., 2016; Saraswat et al., 2016). In addition, severe climatic deterioration has often been regarded as the determining factor leading to the collapse of several ancient agriculturally-based civilizations (Hsu, 1998; Weiss and

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Bradley, 2001). Clearly, detailed spatiotemporal information concerning abrupt and widespread climate change is important for improving our understanding of climate forcing and human responses during the Holocene, and for appraising the possible consequences of extreme climate events under a global warming scenario.

The mid-Holocene environmental transition has attracted much attention from climate scientists and archaeologists, especially Holocene event 3 (HE3, ~4.2 ka), as termed by Bond et al. (1997), because it marks the termination of the Holocene climatic optimum (Perry and Hsu, 2000) and the initiation of the Neoglacial (Solomina et al., 2015). Existing records reveal that ocean surface temperatures decreased by ~1–2 °C during HE3 (Bond et al., 1997; deMenocal et al., 2000b), which persisted for ~300–600 years (Cullen et al., 2000; Perry and Hsu, 2000); while a total duration of up to ~1500 years was recorded in the North Atlantic (Bond et al., 1997, 2001). In addition, HE3 was punctuated by a series of geologically-rapid global cooling and/or dry events (Morrill et al., 2003; Marchant and Hooghiemstra, 2004; Booth et al., 2005; Shanahan et al., 2015) which were superimposed on the gradual drying trend of the mid-Holocene (Morrill et al., 2003; Mayewski et al., 2004; Wanner et al., 2008, 2011; Roberts et al., 2011). Associated with HE3 were the collapse of cultures in Pakistan (Staubwasser et al., 2003; Madella and Fuller, 2006; Macdonald, 2011; Giosan et al., 2012; Ponton et al., 2012; Leipe et al., 2014; Menzel et al., 2014; Prasad et al., 2014a), Mesopotamia (Weiss et al., 1993; Cullen et al., 2000; deMenocal, 2001), China (Jin and Liu, 2002; Wu and Liu, 2004; An et al., 2005; Innes et al., 2014; Zeng et al., 2016; Zhu et al., 2017) and Egypt (Thompson et al., 2002; Marshall et al., 2011; Phillipps et al., 2012).

In high latitudes of the Northern Hemisphere, a peak in detrital carbonate flux on the East Greenland Shelf at 4.7 ka signaled both the beginning of the Neoglacial and a southward expansion of the Arctic sea ice (Jennings et al., 2002). In Europe, a 4.2 ka drought event is recorded by multi-proxy data from a cave flowstone in Italy (Drysdale et al., 2006); diatom assemblages from Montcortès Lake in the Iberian Peninsula indicate that lake levels were lower during a pronounced dry interval from 2360 to 1850 BCE (Scussolini et al., 2011); a decrease in deciduous *Quercus* and *Pinus pinea*-type percentages in Southwest Iberia at ~4.2 ka suggests an abrupt shift to dry conditions (Lillios et al., 2016); and a synthesis of records from the Mediterranean reveals an unusually dry interval from 4.5 to 3.9 ka (Mercuri et al., 2011; Roberts et al., 2011). Evidence from eastern tropical Africa indicates a shift to drier conditions at ~4.0 ka (Marchant and Hooghiemstra, 2004), although at this time wetter conditions were maintained in West Africa (Russell et al., 2003) and in parts of South America (Marchant and Hooghiemstra, 2004); and magnetic and geochemical data from the Holocene sediments of Lake Tana in northwest Ethiopia confirm that the driest interval occurred at ~4.2 ka (Marshall et al., 2011), which is also identified in the Mount Kilimanjaro ice core (Thompson et al., 2002) and in the Mauritian lowlands (de Boer et al., 2014). In eastern Russia, evidence of a cold spell between 4.5 ka and 3.5 ka is provided by a multi-proxy record from Two-Yurts Lake (Hoff et al., 2015). A severe centennial-scale megadrought in mid-continental North America occurred between 4.1 and 4.3 ka (Booth et al., 2005).

In Asia, a record of the concentration of eolian minerals in core M5-422 from the Gulf of Oman reveals an abrupt increase in the content of eolian dolomite and $CaCO_3$ at ~4.1 ka, indicating the onset of dry conditions in Mesopotamia (Cullen et al., 2000); a planktonic oxygen isotope record from core 63 KA from the Arabian Sea shows a shift to heavier $\delta^{18}O$ values at ~4.2 ka, indicating dry conditions in the Indus valley (Staubwasser et al., 2003); multiple lines of evidence from Lonar Lake indicate that the driest conditions in central India occurred from 4.8 to 4.0 ka (Prasad et al., 2014b; Sarkar et al., 2015), in agreement with records from Wadhwana Lake of very dry conditions at ~4.3 ka, (Prasad et al., 2014a); a Holocene pollen record from the northwestern Himalayan lake Tso Moriri suggests that the prolonged Holocene trend towards aridity was punctuated by an interval of increased dryness between ~4.5 ka and 4.3 ka (Leipe et al., 2014); after ~5 ka, the forest vegetation in the Mongolian Altai began to give way to a predominance of open vegetation types as the precipitation amount decreased from 450 to 550 mm/yr to 250–300 mm/yr (Rudaya et al., 2009); the stable oxygen isotopic composition of authigenic lacustrine calcite from Xingyun Lake, in Yunnan Province in China, recorded a substantial positive shift from 5.0 to 4.3 ka, which is coincident with aridity in India and the Tibetan Plateau (Hillman et al., 2017); vegetation history during the Baodun period (~4.5–3.7 ka) in the Chengdu Plain reveals that the dry and cool climate conditions occurred at this period (Zeng et al., 2016); an abrupt climatic transition from wet to dry conditions at ~4.1 ka in the western part of the Chinese Loess Plateau (An et al., 2005), and an exceptional cold event at 4.6–4.3 ka in Hebei Province (Jin and Liu, 2002), indicate short-duration cold and dry events during the mid-Holocene in northern China; at ~4.2 ka there was an increase in cold-tolerant trees at Taihu Lake in the Yangtze coastal plain (Innes et al., 2014) and the climate in the middle reaches of the Yangtze River became rather dry and cold (Zhu et al., 2017); and a rapid and catastrophic fall of the groundwater table of up to ~30 m in the Otindag Desert at ~4.2 ka marks the onset of irreversible desertification in northern China (Yang et al., 2015).

Furthermore, several comprehensive reviews at a regional or global scale of Holocene glacier fluctuations (Solomina et al., 2015) and Holocene climate changes (An et al., 2006; Wanner et al., 2008, 2011; Rudaya et al., 2009; Zhao et al., 2009; Zhao and Yu, 2012; Wang and Feng, 2013) suggest that various climatic proxies recorded a cool and/or dry event in middle to low latitudes of Eurasia at ~5–4 ka. In addition, the wet status during this interval was confirmed in high latitudes of Eurasia (e.g., northern Xinjiang, the northern Mongolian Plateau, the Lake Baikal region and western Russia) as evaporation decreased (Rudaya et al., 2009).

Collectively, these cool and/or dry events in monsoon-influenced regions demonstrate that from 5 to 4 ka a significant large-scale drought in middle to low latitudes of the Northern Hemisphere was a primary response to an interval of weak monsoon strength. The consensus indicates that the mid-Holocene weak monsoon likely resulted from interacting ocean-atmosphere processes in the western tropical Pacific and the equatorial Indian Ocean, such as the El Niño Southern Oscillation (ENSO) (Moy et al., 2002; Marchant and Hooghiemstra, 2004; Macdonald, 2011) and the Indian Ocean Dipole events (Abram et al., 2007; de Boer et al., 2014); and from the southward migration of the Intertropical Convergence Zone (ITCZ, Haug et al., 2001; Wanner et al., 2008; Abram et al., 2009; Sarkar et al., 2015; Shanahan et al., 2015) or cold climatic conditions at high latitudes of the Northern Hemisphere (Bond et al., 1997; Menzel et al., 2014; Fan et al., 2016). In addition, a regional interpretation considered this to be an irreversible region-wide hydrologic event (Yang et al., 2015). Ecosystem degradation was one of the most important effects of HE3, resulting in the abandonment of dwellings and the migration of Neolithic peoples to more habitable areas (Hsu, 1998; Perry and Hsu, 2000; Giosan et al., 2012; Jia et al., 2016), and the forced modification of agricultural systems such as the increased cultivation of drought-tolerant crops like millet (Giosan et al., 2012) and the spread of wheat in East Asia (Dodson et al., 2013; Chen et al., 2015a). Giosan et al. (2012) suggested that the effects of HE3 caused a decrease in the total settled area and in settlement sizes in the Indus Valley since 3.9 ka. In addition, the combination of decreased agricultural productivity and increased population in habitable areas triggered