



Seasonal evolution differences of east Asian summer monsoon precipitation between Bølling-Allerød and younger Dryas periods

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Abstract

The rapid warming trend during the last deglaciation triggered significant global climate instabilities due to a complex non-linear response of the climate system to the gradual increase in insolation over the northern hemisphere. Although climate impacts can be detected globally, major regional imprints such as seasonal evolution and changes of the East Asian Summer Monsoon (EASM) during the last deglaciation are still poorly constrained due to a lack of comprehensive proxy data. In this study, we compare how the extreme climate shifts are linked to changes in EASM precipitation over China between the unusually warm Bølling-Allerød (BA) interstadial and the following strong cooling of the Younger Dryas (YD) stadial. Our analysis is based on the transient atmosphere-ocean simulations of TraCE-21ka, in addition to new results from high-resolution simulations of the CESM1 model for the BA and YD time slices. We find that the earlier onset and stronger intensity of the EASM in the BA interstadial lead to more precipitation in early summer (May–June) but drier conditions during mid-summer (July–August) over Southern China compared to a stadial climate during the YD episode. For Northern China, we find the opposite response. The insolation change in spring and the forced response of the atmospheric system are thought to be responsible for these differences. Relative to the YD episode, the hemispheric temperature gradient during the BA period is enhanced due to the asymmetric warming between the two hemispheres, leading to an intensified northward equatorial cross flow. Combined with a stronger sensible heating of the Tibetan Plateau in spring and the related earlier northward shift of the westerly jet, the early onset of the EASM is triggered. The latent heat release, which is accompanied by the onset of the EASM and the sudden increasing precipitation over Southern China in early summer, contributes to the westward shift of the Western Pacific Subtropical High (WPSH) and eastward movement of the South Asia High (SAH) in mid-summer. Under the above conditions, Southern China experiences a hot and dry climate, while Northern China receives more precipitation. Additionally, the La Niña-like pattern of the equatorial Pacific also partly contributes to the strong EASM in the warm period by influencing the WPSH location and Pacific-North American (PNA) teleconnection pattern.

Keywords EASM precipitation · Seasonal evolution · BA period · YD period

1 Introduction

The last deglaciation, spanning from ~ 19 to 11 ka BP from the late Last Glacial Maximum (LGM) to the early Holocene, is an unstable period with a series of abrupt climate shifts superimposed on the increasing trend of temperature due to a complex non-linear response of the climate system to the gradual increase in insolation in the northern hemisphere (Karpuz & Jansen, 1992; Mayewski & Bender, 1995; Alley & Clark, 1999; Denton et al., 2010). The well-known abrupt climate events such as the cold Heinrich Stadial (HS1), the warm Bølling-Allerød (BA) interstadial, and Younger Dryas (YD) stadial have been originally discovered from deep sea sediments in the North Atlantic and ice cores from Greenland (Heinrich, 1988; Dansgaard et al., 1989; Broecker et al., 1992; Bond et al., 1997). These abrupt climate events reflect the external forcing changes and related feedbacks or adjustments of the climate system. The warming in the northern high latitudes during the last deglaciation has been found to be closely related to the increasing summer insolation and a resumption of northward oceanic heat transport in response to the formation of North Atlantic deep water (NADW) after the HS1 event (Lehman & Keigwin, 1992). Furthermore, a regional to global synchronization of the deglaciation is caused by the ice-albedo, sea level, and greenhouse gas feedbacks, together with tropical warming in response to polar amplification (Alley & Clark, 1999). The propagation of the climate signal has been further verified by massive proxy evidence in the Asian region (An et al., 1993; Guo et al., 1996; Wang et al., 2001; Yuan et al., 2004; Sun et al., 2011). The deglacial meltwater pulse in the North Atlantic region could induce the YD event and prompt climate cooling in East Asian monsoon regions (Nakagawa et al., 2006; Renssen et al., 2015). The loess is a manifestation of the European YD period, and the grain-size maxima reflects the climate linkage of climate between North Atlantic and China under the effect of the westerlies (An et al., 1993; Porter & An, 1995; Qiang et al., 2004). In addition, the swamp sediment sequences and isotope signals of stalagmites clearly reflect the instability during the last deglaciation, suggesting a large-scale coherence of monsoonal precipitation and the temperature in northern high latitudes (Zhou et al., 1999; Wang et al., 2001; Liu et al., 2008; Shinozaki et al., 2011).

The teleconnection between the Asian monsoon and the climate in high latitudes reflects the non-linear response of atmospheric circulation to orbital forcing and the coupling processes between different components of the global system (An, 2000; Yang et al., 2010). The Atlantic meridional overturning circulation (AMOC) is a primary driver for the abrupt climate events, and the westerlies play a crucial role in transmitting this signal from the North Atlantic to the Asian regions (Guo et al., 1996; Zhou et al., 1999; Nakagawa et al., 2006; Shen et al., 2010; Clark et al., 2012). The winter monsoon, especially cold surges, acts as the conveyor in the process of transmitting the temperature anomaly from the Arctic to the tropics (Chu et al., 2017). Additionally, the changes in sea ice cover might influence the East Asian monsoon variability through indirectly generating the occurrence of ENSO events (An, 2000; Wang et al., 2001; Qiang et al., 2004).

The monsoon is the dominant seasonal mode of climate variability over the Asia region and plays the leading role in the process of moisture and energy transportation. It comprehensively reflects the interactions of the climate subsystems and their combined effect within the boundary conditions imposed by the East Asian continent and solar radiation (An, 2000). The evolution of the Asian monsoon has been the focus of scientific research owing to its complexity and importance. Nevertheless, our understanding of EASM evolution during the last deglaciation is mainly based on loess-paleosol sequences and reconstructions of

stalagmites from several caves and is still insufficient and debated so far due to uncertainty and locality of the proxy data. The boreal summer insolation change is thought to be the primary factor for the EASM evolution, but it cannot explain the substantial regional character of the monsoon precipitation (Peterse et al., 2011; Mohtadi et al., 2016). The proxy data from stalagmites are determined by the East Asian hydrological environment, which is closely related to the migration of the EASM-related rain belt that is determined by the intensity and location of both the Intertropical Convergence Zone and the Western Pacific Subtropical High (Shinozaki et al., 2011). The different meridional responses of precipitation in the EASM area between BA and YD periods indicate that other climate factors beyond the North Atlantic climate jointly modulated the EASM, highlighting the complex dynamics of the East Asian monsoon system and challenges for future projections (Dong et al., 2015; Ding et al., 2016; Wen et al., 2016; Wang et al., 2016).

As above mentioned, although climate impacts can be detected globally, major regional imprints during the last deglaciation are still poorly constrained due to lack of comprehensive proxy data. The response of the EASM under various abrupt climate events and its physical mechanism is an important and meaningful issue but cannot be well addressed only based on proxies which are limited by low temporal resolution and restricted spatial coverage. To overcome these limitations, full complexity (paleo-)climate model simulations are ideal tools to gain spatiotemporal climate reconstructions and study driving mechanisms and climate impacts in a physical-dynamical consistent way. For the last deglaciation, we make use of the total forcing numerical experiments of the climate evolution during the last 21 ka (TraCE-21ka). The TraCE-21ka simulation successfully reproduces the abrupt BA warming as a transient response of the Atlantic meridional overturning circulation (AMOC) to a sudden termination of freshwater discharge to the North Atlantic Ocean at the end of HS1 (Liu et al., 2009). The YD stadial in TraCE-21ka was realistically simulated through testing different freshwater hosing events at the end of the Allerød (He, 2011; He et al., 2013). In addition, new results from high-resolution simulations of the CESM1 model for the BA and YD time slices are investigated (Schenk et al., 2018). Combining these simulations with previous proxy data results, we will discuss the seasonal evolution difference of the EASM precipitation between the warm and cold periods during the last deglaciation. We further reveal the related physical mechanism from the perspective of the response of the atmospheric system to external forcing change in this study.

2 Data and experiments

We make use of two climate model datasets in this study. One is derived from the simulation of the transient climate of the last 21,000 years (TraCE-21ka) conducted by the National Center for Atmospheric Research (NCAR) using the coupled climate model CCSM3 (Liu et al., 2009; He, 2011; He et al., 2013). The experiment is based on a fully coupled atmosphere-ocean model version. The coupled model was run at a relatively coarse T31_gx3 grid resolution with around $3.75^\circ \times 3.75^\circ$ (~ 375 km) horizontal resolution for the atmosphere with 26 vertical levels. The model is forced with transient greenhouse gas concentrations (Joos & Spahni, 2008) and orbit-driven insolation changes. Transient boundary conditions include the ICE-5G ice sheets and topography (Peltier, 2004). The simulation starts at 22,000 years BP and finishes in 1990 AD, and the decadal mean for each season is used in this study. For a detailed description of the model simulations, see He (2011).

The other used simulation experiment consists of steady-state simulations using a high-resolution paleoclimate version of $0.9^\circ \times 1.25^\circ$ (~ 100 km) of the CESM1.0.5 model for the YD period (12.17 ka BP), Bølling (14.3 ka BP), Allerød (13 ka BP), and early Holocene (11 ka BP). The high-resolution experiments are based on a coupled atmosphere-land-ice component set of CESM1 (F1850) where SST and ice fraction for each time period are prescribed from the TraCE-21ka results as 100-year climatology around each time slice. Greenhouse gas concentrations and orbital forcing are the same 100-year averages as SST and are identical to the TraCE-21ka values for each time period. The simulations are performed using modified realistic boundary and forcing conditions taking into account a new reconstruction of continental ice sheets (GLAC-1B) and corresponding changes in land-sea distribution representing low sea level stands. All CESM1 experiments analyzed here follow the same setup as described for the YD and AL experiments (Schenk et al. 2018). Each experiment is integrated for 180 years, and the monthly data in the last 30 years are used for analysis in this study.

3 The simulated climate during the last deglaciation

3.1 Variations of the global surface temperature and precipitation

As mentioned, the last deglaciation is featured by climate instability with several abrupt events superimposed on the rising temperature trend. Can the model successfully reproduce the climate characteristics of the last deglaciation revealed by robust proxy-evidence? Figure 1a presents the variation of the surface temperature since the Last Glacial Maximum derived from the TraCE-21ka results. Apparently, the warming trend is dominant in the global climate during the past 21,000 years. We also notice that different from the stability from 10 ka BP to the present, the instability is mostly a feature of the deglaciation from 18 to 10 ka BP. Several events such as cold event H1, BA warm period, and YD cold period can be clearly observed. The maximum temperature in Bølling warm period approximately occurred at 14.3 ka BP with about 2°C magnitude of abrupt warming relative to H1. After a short cold interval at 14 ka BP (Older Dryas, OD), the temperature rises again at 13 ka BP (the Allerød warm period) but with moderate warming in comparison to the Bølling episode. The simulated YD period, which is thought to be caused by a slowdown of the Atlantic meridional overturning circulation by meltwater discharges into the North Atlantic (Renssen et al., 2015), occurred from ~ 12.7 to ~ 11.7 ka BP with the coldest simulated period around 12.17 ka BP with a global temperature drop of 0.5°C . However, rather than a homogeneous global cooling, die AMOC disturbance leads to notable

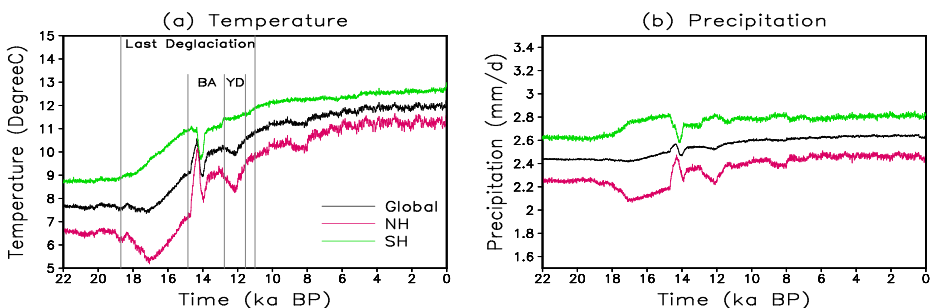


Fig. 1 The variations of annual surface temperature (a) and precipitation (b) derived from the TraCE-21ka results

hemispheric climate differences. The temperature of the Southern Hemisphere (SH) does not show any sign of decline during the period of H1 that occurred in the Northern Hemisphere (NH) at about 17 ka BP. This is in agreement with that the SH warming is about 3 ka prior to the NH (Barker et al., 2009; Ayliffe et al., 2013; Xu et al., 2013). Similarly, the YD period, which well documented by proxies in mid- to high latitudes of the NH (Björck 2007; Denton et al., 2010), showed little evidence in its SH counterpart, implying that the cold events during the last deglaciation might not be a dominant global phenomenon (Barker et al., 2009).

Relative to temperature, the precipitation variation is more stable with evident anti-phase change between two hemispheres. From the zonal mean section of the temperature and precipitation (Fig. 2), the differences between various latitudes are notable, especially during the period from 15 to 11 ka BP, indicating that the meridional inconsistency of precipitation variation might be a response and adjustment to the climate instability of the last deglaciation.

3.2 Stadial-interstadial differences of EASM precipitation

Based on testing an array of freshwater hosing experiments conducted for TraCE-21ka (He 2011), the climate instability of the last deglaciation has been successfully captured in the simulation. It is the basis for us to further investigate how the atmospheric system responds to external forcing under the unstable climate background and its influences on the EASM precipitation pattern. Taking the reconstruction and simulation into consideration, three warm periods (interstadials) and one cold episode (stadial) including 14.3 ka BP for the Bølling period, 13 ka BP for Allerød period, 11 ka BP for early Holocene, and the 12.17 ka BP for YD cold period are selected for further examining their reactions to climate instability. The positive anomalies in temperature between 11 and 12.17 ka BP nearly dominate the global scope with the amplitude exceeding 3 °C in the high latitudes of the NH and polar region of the SH (Fig. 3). Differently, the positive and negative difference values between the BA and the YD period are, respectively, distributed in NH and SH. It means that the warming in the BA period is asymmetrical and regional, which might lead to the precipitation response through influencing the atmospheric system.

Since the EASM precipitation is the focus of this study, we present the precipitation distributions of the four periods for only that region. The precipitation pattern is heavily affected by the topography elevation and land-sea distribution which is closely related to the

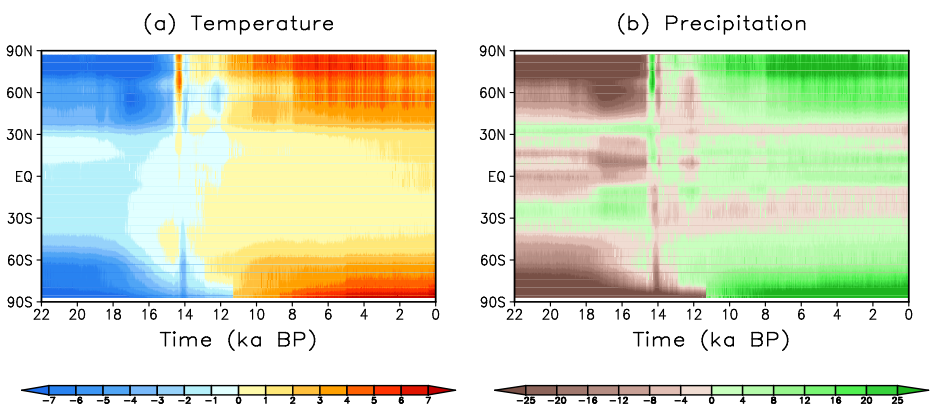


Fig. 2 The zonal mean of annual surface temperature (a with units in °C) and precipitation (b with units in mm d⁻¹) anomaly in respect to the whole period derived from the TraCE-21ka results

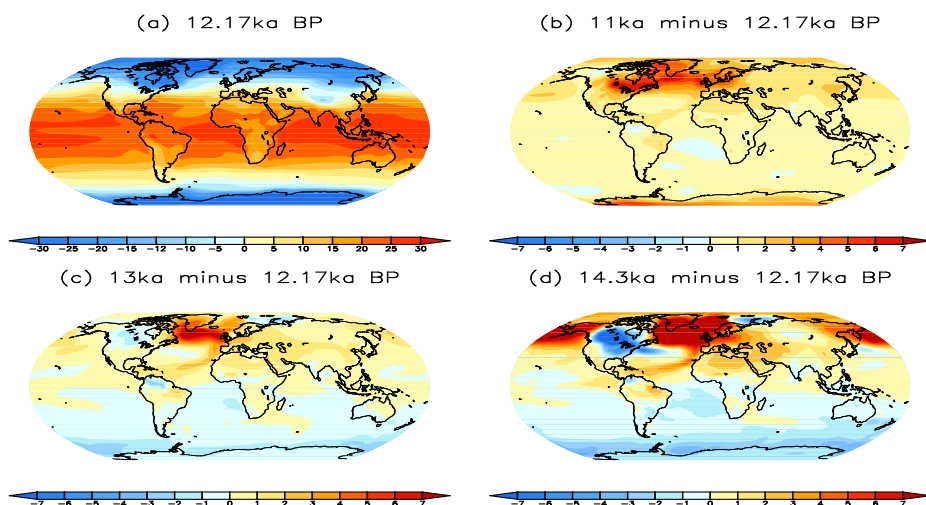


Fig. 3 The annual surface temperature distribution during 12.17 ka BP (a) and differences between the warm periods of 11 ka (b), 13 ka (c), 14.3 ka BP (d), and the cold period of 12.17 ka BP derived from the CESM high-resolution simulation (units in °C)

model resolution. Due to the complexity of East Asian terrain, the coarse resolution results of the TraCE-21ka might not meet our requirement for detailed regional studying the EASM precipitation distribution. The continental precipitation maximum from TraCE-21ka is located at the upper reaches of Yangtze River, east to the Tibetan Plateau, which has been verified as an unrealistic center of precipitation partly due to the inaccurate handling of terrain in the coarse resolution model (Yu et al., 2000). In contrast, the precipitation pattern in the high-resolution simulation, which is characterized by more precipitation in Southern China and less in the Northern China, is in accordance with the modern observation and more reasonable; therefore, the outputs of the high-resolution CESM1 model results are used for the following reanalysis (figures not shown here).

The summer mean precipitation (May to September) over East Asia in 12.17 ka BP shown in Fig. 4 clearly reflects the characteristics of the EASM with decreasing precipitation from the southeast to the northwest of China. In comparison to the cold YD stadial, a common feature of the three warm interstadials is increased precipitation in western China, Hetao and Huang-huai River Basin and decreased precipitation in South China and parts of northeast China. It is generally accepted that the water vapor source for the EASM precipitation mostly depends on the southerly transportation from the South China Sea and western Pacific (Li, 1999; Simmonds et al., 1999; Zhang, 2001; Sun et al., 2011;). North China will be wetter with increased precipitation if the EASM is strengthened, while it will be drier under the weakened EASM condition. Therefore, it can be deduced that the EASM during the warm periods during the deglaciation is stronger than in YD, which is in accordance with the results revealed in stalagmite, loess, and lake sediment records (Wang et al., 2000; Qiang et al., 2004; Hu et al., 2008; Shen et al., 2010; Liu et al., 2014a; Orland et al., 2015; Chen et al., 2015).

The onset time of the EASM is the key point for its northward extension and precipitation distribution (Krishnamurti & Ramanathan, 1982; Wang et al., 2004; Zhang et al., 2004; Zhou et al., 2007; Ha et al., 2018). The onset of the EASM is generally at mid-late May with a sudden rise of precipitation in South China and occurrence of southwesterly flow over the South China Sea as shown in the Fig. 5a. Accompanied by the first northward jump of the

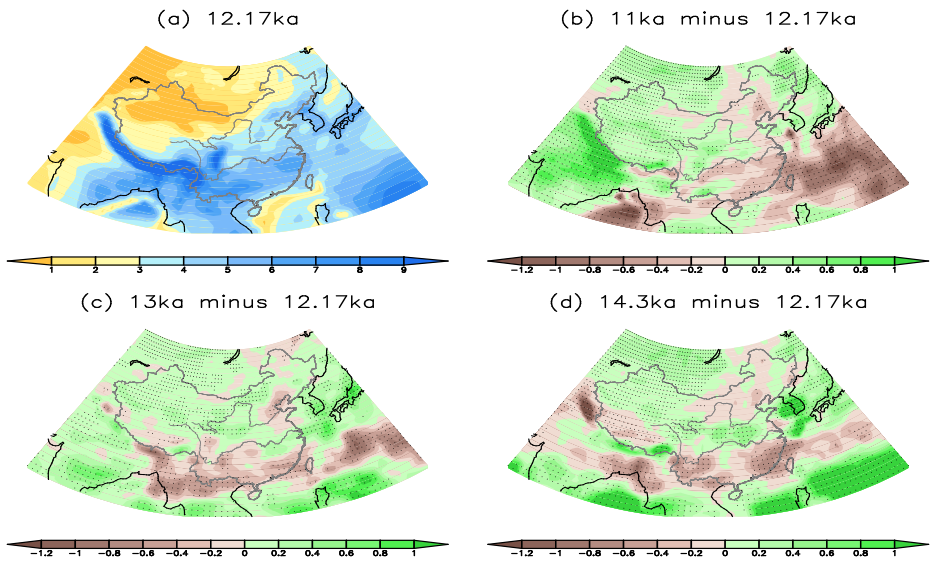


Fig. 4 The distribution of precipitation averaged from May to September during 12.17 ka BP (a) and differences between the warm periods of 11 ka (b), 13 ka (c), 14.3 ka BP (d), and the cold period of 12.17 ka BP derived from the CESM high-resolution simulation (units in mm d^{-1}) (The black dots are for the areas exceeding 95% significance level)

western Pacific subtropical high (WPSH) in June, the rain belt moves northward to the Jiang-huai River Basin from South China (corresponding to the “Meiyu” season). Following the second jump of the WPSH in early July, the rain belt shifts further to North China and brings precipitation to the semi-arid area, while the climate in South China is intolerable with intense temperature and humidity under the control of the WPSH. The EASM retreats southward very quickly in September, and the climate in most areas of China is pleasant under the continental high pressure. It has been revealed that the early onset of the EASM corresponds to its strong intensity, far northward extension, and more precipitation over North China, while the late onset and weakened intensity of the EASM tends to cause more rainfall concentrated over South China (Zhang et al., 2004; Zhou & Chan, 2007; Shen et al., 2010; An et al., 2012; Wen et al., 2016; Kelly & Passchier, 2018).

The seasonal evolution differences of EASM precipitation averaged over 105–122°E between BA warm and YD cold periods are characterized by positive values during May–June and negative values during July–August in the latitudes of 25–35°N (Fig. 5). The situation in the 35–45°N region is inverse. The precipitation distribution indicates that the onset of the EASM during the BA period is earlier than that during the YD period, corresponding to an intensified EASM and earlier northward shift of the rain belt. This explains the increased precipitation in Southern China in early summer and wet climate in the Northern China in mid-summer during the warm period. The simulation results are in accordance with the proxy data (An et al., 1993; An, 2000) but further reveal the seasonal evolution differences of EASM precipitation between two periods. This raises the question: What factors should be responsible for this seasonal evolution difference of the EASM precipitation between two climate events during the last deglaciation? The following section will discuss the possible reasons from the prospective of the response of the atmospheric system to external forcing.

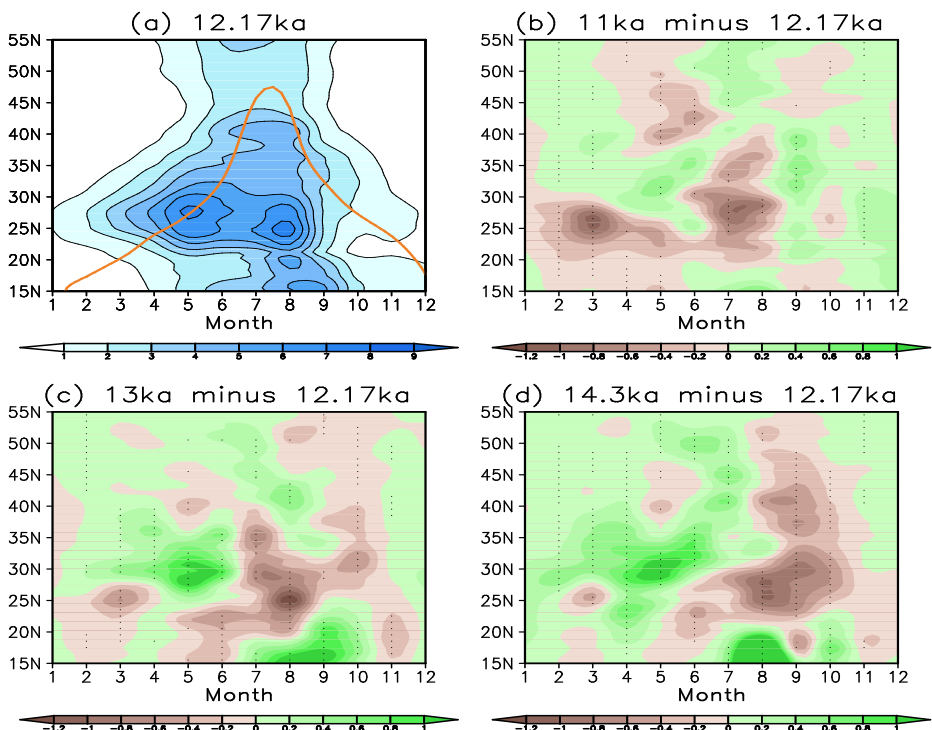


Fig. 5 The latitude-seasonal variation of precipitation averaged of 105–122°E during 12.17 ka BP ((a) the orange curve is 5800 gpm contour at 500 hPa level that could represent the seasonal shift of the WPSH) and differences between the warm periods of 11 ka (b), 13 ka (c), 14.3 ka BP (d), and the cold period of 12.17 ka BP derived from the CESM high-resolution simulation. (Units in mm/d; The black dots are for the areas exceeding 95% significance level)

Considering the circulation similarity of the Bølling and Allerød period, the 13 ka BP is taken as a case for discussing the physical mechanism between BA period and YD episode.

4 Possible reasons for EASM precipitation differences between BA and YD

4.1 Insolation changes in spring

Figure 6 gives the variations of the simulated surface temperature (TS), net shortwave radiation at the top of the atmosphere (FSNT), and insolation averaged over the NH. It is noted that the FSNT curve matches the insolation well in each season but further reflects the energy balance of the earth system under the external forcing. Therefore, the abrupt fluctuations can be found in FSNT but not existed in insolation variation. The insolation variation due to earth's orbit relative to the sun is a key factor during the last deglaciation and contributes to abrupt changes of many environmental indices. The high correlation of 0.99 between the decadal mean surface temperature and simulated FSNT averaged of the NH further verifies this point (Fig. 6a). The summer insolation is usually used for explaining the climate change derived from the proxy data (Mayewski & Bender, 1995; Alley & Clark, 1999; Clark et al., 2012; Chen et al., 2015; Renssen et al., 2015). However, the seasonal

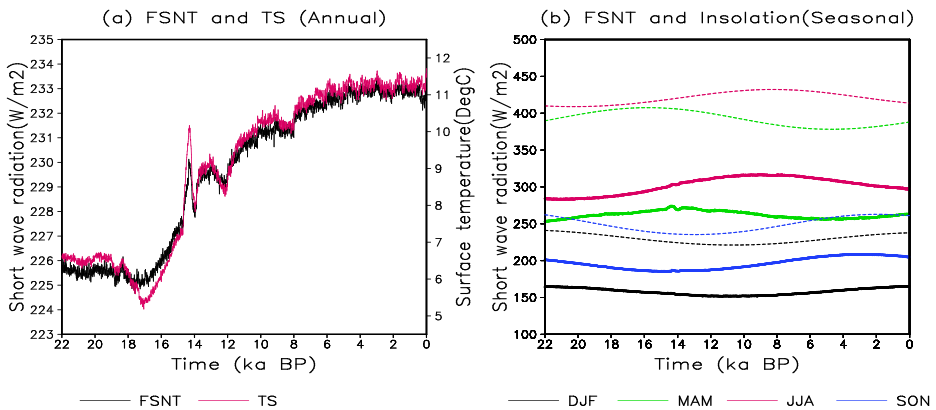


Fig. 6 The variations of annual surface temperature (TS of red curve in **a**), flux of net shortwave radiation at the top of the atmosphere (FSNT of black curve in **a** and solid curves in **b**) and insolation (dashed curves in **b** with same color as FSNT in each season) averaged of the Northern Hemisphere derived from the TraCE-21ka results

transition of EASM precipitation cannot be solely interpreted through summer insolation changes. The correlation between the annual surface temperature and the spring FSNT from 15 to 11 ka BP reaches 0.88, higher than with other seasons. Particularly, there is a peak in the BA period and a valley in the YD period in the variation of the FSNT series, which cannot be observed in other seasons. This phenomenon might be caused by the non-linear response of climate system to a sudden termination of freshwater discharge to the North Atlantic Ocean in the model that is significantly reflected in the spring season and need further study. Thus, the consistency of the temperature and the spring FSNT series implies that the spring insolation might play more crucial role on the climate variation of the last deglaciation than other seasons. As the onset, intensity, and northward extend of the EASM are closely related to the preceding thermal condition of the climate system, it is no doubt that the spring solar radiation exert profound influences on the timing and evolution of the EASM.

4.2 Atmospheric response to the abrupt climate transition

4.2.1 Strengthened cross-equatorial flow caused by inter-hemispheric temperature contrast

The cross-equatorial flow is the fundamental component of the EASM system as a conveyor of water vapor between the NH and SH. In the boreal summer, the temperature in the SH is colder than in the NH; as a result, the air pressure of the SH is higher than that in the NH. Therefore, the hemispheric air pressure gradient force causes the air flow from the SH passing across the equator to the NH. The westerly flow from the Indian Ocean originally coming from the Somali jet, the southwesterly flow over the Bengal Bay and South China Sea, and the southeasterly flow along the edge of the WPSH are the main paths of water vapor transportation for the EASM precipitation (Li, 1999; Zhang, 2001; Zhou & Yu, 2005; Sun et al., 2011). Previous research also pointed out that the onset and the intensity of the EASM are closely connected with the strength of the cross-equatorial flow (Zhang et al., 2004). A key question for understanding EASM changes under different climate states is how strongly the seasonally varying temperature contrast of both hemispheres affects the timing and strength of monsoon onset. Partly related to the timing, it is crucial to understand how far north/south the monsoon reaches over the continents which determines the general trends to wetter or drier conditions.

The onset of the EASM normally occurs in mid-late May and is signaled by the occurrence of southwesterly winds over the South China Sea (He & Liu, 2016). The onset timing of the EASM is largely determined by the intensity of southwesterly flow. From Fig. 7a, the hemispheric temperature differences reverse in May from a colder to a warmer NH relative to the SH. The hemispheric temperature difference reverses again in October, implying that the NH is warmer than SH from May to September corresponding to the onset and retreat of the EASM. Generally, an earlier onset of the EASM corresponds to the stronger EASM, and vice versa. The hemispheric temperature difference in the BA warm period is about 1–2 °C higher than that in YD cold period throughout the year as demonstrated in Fig. 7a. It means that the amplitude of the hemispheric temperature difference of the BA period is larger in the summer half year and smaller in the winter half year in comparison with that of the YD period. Consequently, the hemispheric sea level pressure takes the contrary evolution except during June and July that may be caused by other factors (Fig. 7b). Figure 8 presents the zonal distribution of the meridional wind along 20°N in May at 850 hPa and 200 hPa level, respectively. The positive value refers to the southerly anomaly, which occupies most area at 850 hPa level, indicating a stronger cross-equatorial flow in the BA period than in the YD period. Thus, the intensified cross-equatorial flow is critical for triggering the earlier onset of the EASM and its more northward shift in the BA period.

4.2.2 Earlier onset of EASM triggered by intensified surface sensible heating of the Tibetan Plateau and the related earlier northward shift of the westerly jet stream

As the largest topographic feature in the world, the Tibetan Plateau, often referred to as the “third pole”, plays an essential role in the formation and evolution of the East Asian monsoon via exerting strong thermo-dynamic influences on the atmospheric circulation (Hsu & Liu, 2003; Duan et al., 2011, 2013; Wu, et al., 2012, 2017; Wang et al., 2014; Son et al., 2019; Zhan, et al., 2019). Particularly, the surface sensible heat flux caused by the strong warming of the Tibetan Plateau in spring generates a thermal cyclonic circulation at low levels and a corresponding anticyclone (the South Asia High) at upper levels of the troposphere. The vertical flow caused by the convergence of the cyclonic circulation at low level acts as a “pump” which not only disrupts the Hadley circulation but also accelerates the cross-equatorial southerly flow. In addition to pumping water vapor from

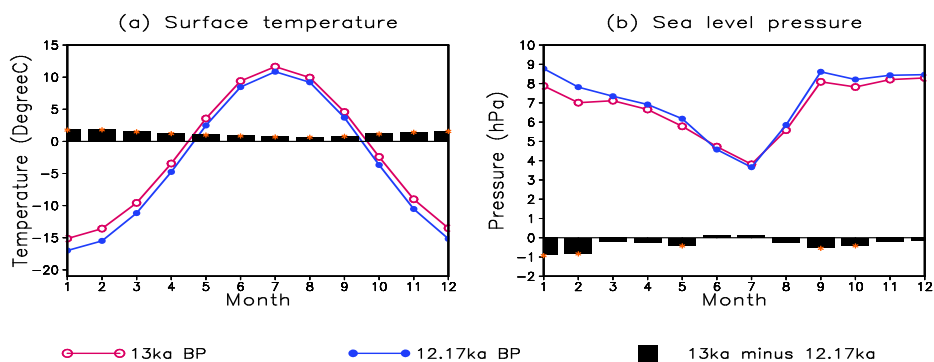


Fig. 7 The annual variation of hemispheric difference (NH minus SH) of surface temperature (a) and sea level pressure (b) derived from the high-resolution simulation (red empty dot line is for 13 ka BP, blue solid dot line for 12.17 ka BP, and black bar for difference between 13 ka and 12.17 ka BP with the orange asterisk for that exceeding 95% significance level))

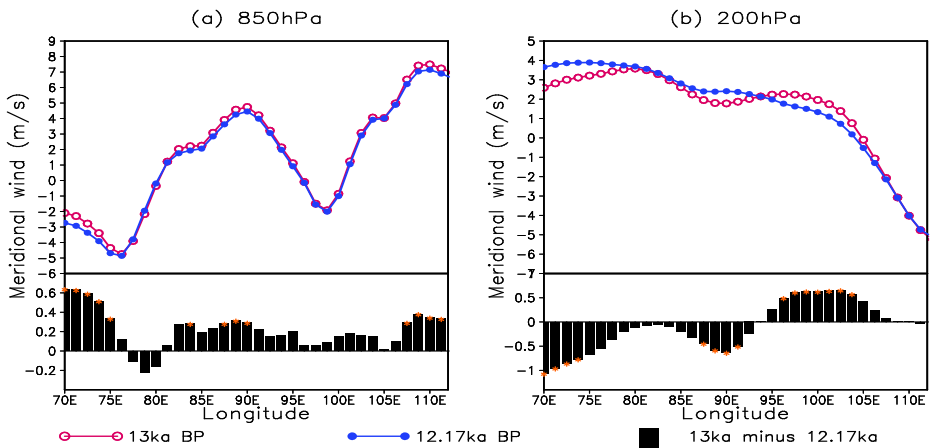


Fig. 8 The zonal distributions of meridional wind along 20°N in May, respectively, at 850 hPa (a) and 200 hPa level (b) derived from the high-resolution simulation (red empty dot line is for 13 ka BP, blue solid dot line for 12.17 ka BP, and black bar for difference between 13 ka and 12.17 ka BP with the orange asterisk for that exceeding 95% significance level)

sea to land and from the lower to the upper troposphere, it also generates a subtropical monsoon-type meridional circulation subject to the angular momentum conservation, providing an ascending-air large-scale background for the development of the Asian summer monsoon (Wu et al., 2017). In addition, the location of East Asian westerly jet stream, which is closely related to the topographical forcing and diabatic heating of the Tibetan Plateau (Zhang et al., 2006; Son et al., 2020), is critical for determining the stepwise transitions in East Asian rainfall seasons (Chiang et al., 2015, 2017). Thus, the thermally driven circulation changes over the Tibetan Plateau are thereby conducive to the onset of the EASM.

The sensible heating anomalies of the Tibetan Plateau between 13 and 12.17 ka BP are presented in Fig. 9a, b. The temperature in the BA period is higher than during the YD period throughout the year. Especially, it is much higher in March and April, indicating that the spring warming of Tibetan Plateau in the BA period is much stronger, which is in accordance with the increasing solar radiation in this season. The sensible heating over the Plateau, which is basically determined by the surface-air temperature differences, is consequently strengthened in BA period and contributes to the early northward shift of the westerly jet in Spring as demonstrated in Fig. 9d and leads to the early onset of the EASM to large extent, which is consistent with the find of Chiang et al. (2015). After the onset of the EASM, the rapid increased rainfall leads to the decline of surface-air temperature and the sensible heating, and the heating of the Tibetan Plateau is mostly replaced by the latent heat flux released by the precipitation. Therefore, it well explains that the negative value of sensible heating differences of Tibetan Plateau between BA and YD period in May observed in Fig. 9b and the intensified and more southward jet stream in summer during BA period in Fig. 9c, d.

4.2.3 Eastward shift of the south Asian high and westward movement of the western Pacific subtropical high

The WPSH and the SAH, respectively, centered at the mid-low level and upper level of troposphere, are the two indispensable centers of action in the EASM system. The center of the

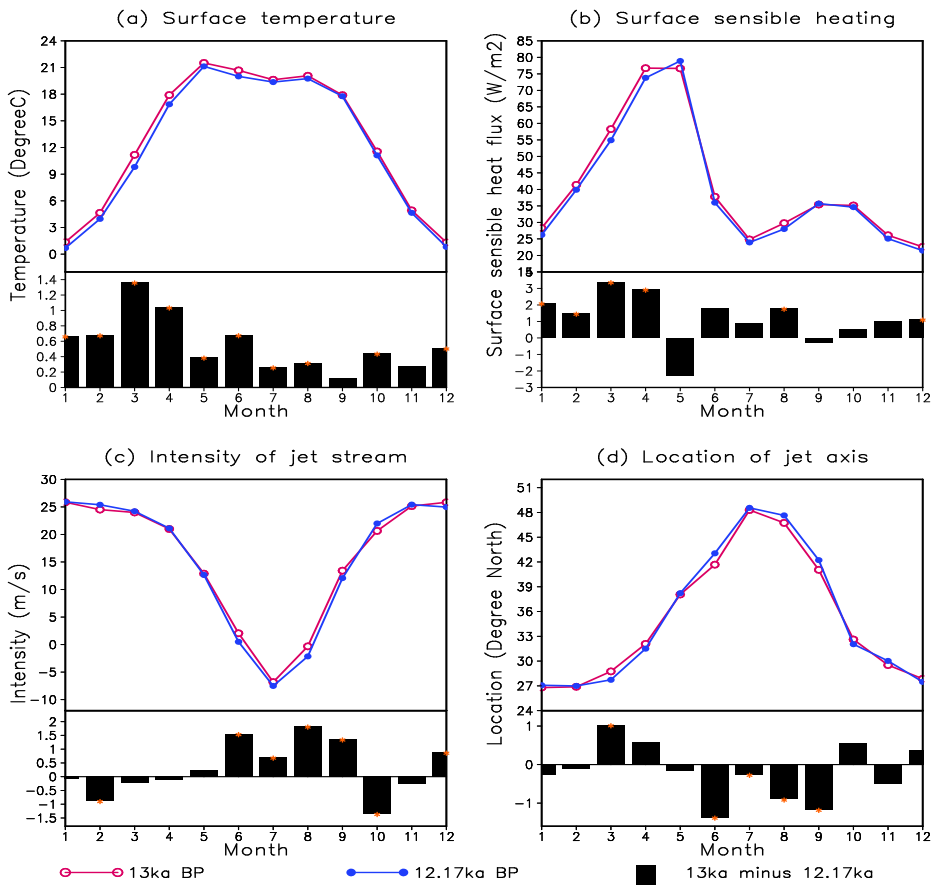


Fig. 9 The annual variations of surface temperature (a), sensible heating (b) of the Tibetan Plateau (averaged of 80–100°E, 20–30°N), intensity (c; averaged zonal wind speed of 70–130°E, 20–50°N) and axis location of the westerly jet stream at 200 hPa level (d; mean latitude of maximum zonal wind of 70–130°E) (Red empty dot line is for 13 ka BP, blue solid dot line for 12.17 ka BP, and black bar for difference between 13 ka and 12.17 ka BP with the orange asterisk for that exceeding 95% significance level)

SAH is generally located at the 100 hPa level due to the strong direct heating of the Tibetan Plateau and the Iran Plateau to the upper troposphere (Zhang et al., 2002). The location and intensity of the SAH, which is crucial for the EASM precipitation, are closely related the heating of the Tibetan Plateau (Zhang et al., 2002; Qian et al., 2002). In addition, the SAH is characterized by its thermal preference; i.e., its center tends to shift toward the heating position (Qian et al., 2002). As previously discussed, the onset of the EASM in the BA period is earlier than YD episode due to the strengthened cross-equatorial flow and the sensible heating of the TP. Namely, the more precipitation in Southern China in May–June leads to much latent heat release in the atmosphere and causes a shift eastward of the SAH according to its thermal preference as demonstrated in Fig. 10 (Wei et al., 2015). The SAH in the BA period is stronger and more eastward in comparison with that in the YD period. Previous studies show that the SAH and WPSH are related to the concept of “come closer or go further” (Qian et al., 2002). It means there are two common patterns for the zonal positions of the two monsoon subsystems: one is “close” and the other is “far away”. In the case of the BA period, the eastward shift of

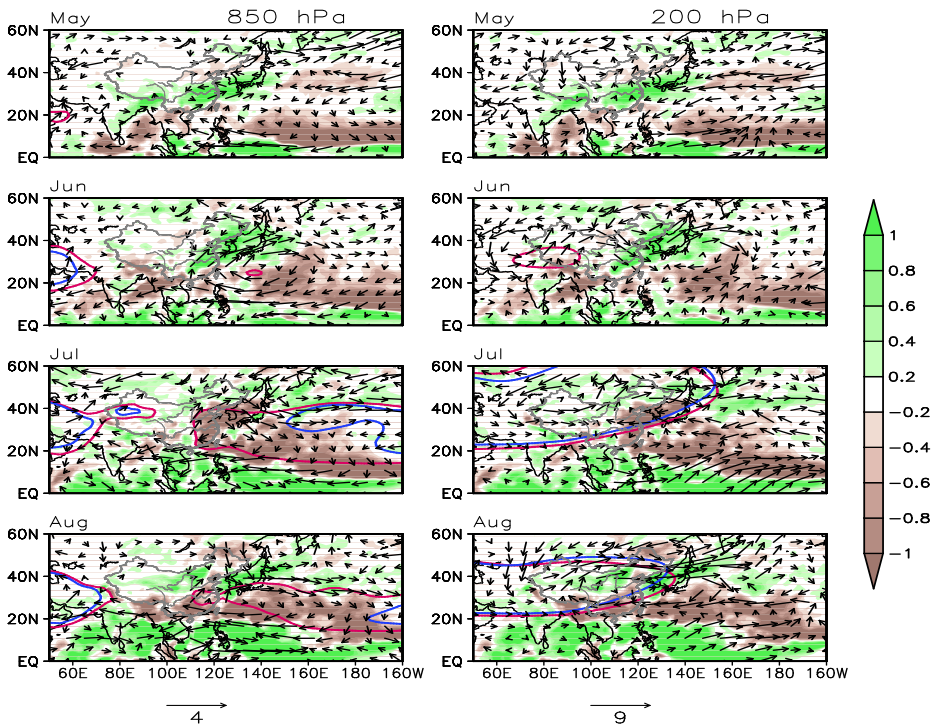


Fig. 10 The differences of precipitation (shading in units of mm/day) and wind (vector in units of m/s) between 13 ka BP and 12.17 ka BP from May to August, respectively, at 850 hPa level (left panel; the red curve is for the western Pacific subtropical high with value of 5850gpm at 500 hPa level in 13 ka BP and the blue is for 12.17 ka BP) and 200 hPa level (right panel; the red curve is for the south Asia high with value of 16500gpm at 100 hPa level in 13 ka BP and the blue is for 12.17 ka BP)

the SAH and the westward movement of the WPSH correspond to the first pattern. The circulation distribution at 850 hPa level further verifies this point. In contrast with the YD period, the WPSH in the BA period is more westward and much stronger, so Southern China receives less precipitation under the control of the WPSH.

Additionally, the EASM is also affected by the thermal condition of the equatorial Pacific. The higher temperature in the warm pool of western Pacific generally corresponds to a strong EASM via exerting the impact on the teleconnection pattern of Pacific-North America (PNA) and the location of the western Pacific subtropical high (Wang et al., 2017). From the SST difference between the two periods (Fig. 11), the negative value is positioned the eastern

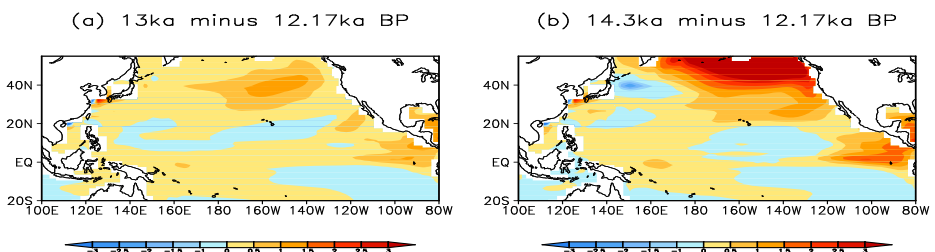


Fig. 11 The sea surface temperature differences between 13 ka (a), 14.3 ka BP (b), and 12.17 ka BP (Units in °C)

Pacific, while the positive value is located at the western Pacific area, which is similar to La Niña pattern. According to a previous study (Huang, et al. 2003, 2006), the La Niña-like pattern are conducive to the strong EASM and a more northward extension of the rain belt in Eastern China through influencing the location and the intensity of the WPSH, which is consistent with our results.

5 Conclusion and discussion

The last deglaciation is a very unstable period due to a strongly non-linear response of the climate system to the gradual increase in orbital summer insolation over the NH. In this study, we compare the seasonal evolution differences of the EASM precipitation between the BA warm period and YD cold period based on climate model simulations. The results show that the earlier onset and stronger intensity of the EASM in the warm interstadial period lead to a shift in the timing and regional pattern with more precipitation in May–June and drier conditions in July–August over Southern China and the opposite phenomenon occurring in Northern China with respect to the YD cold period. As the boreal spring insolation is rapidly rising, the warming in the NH is higher than in the SH. As a result, the hemispheric temperature gradient forcing is enlarged due to the uneven warming amplitude, leading to the stronger equatorial cross flow from S to N in late spring, which is the original source of the EASM. In the same time, the sensible heating of the Tibetan Plateau is enhanced due to the increased spring insolation and leads to a stronger thermal low and horizontal advection of air. Combining the stronger cross equatorial flow and the intensified heating of the Tibetan Plateau as well as the earlier northward shift of westerly jet, the EASM onset is triggered earlier in BA than YD. Accompanied by the onset of the EASM, more water vapor is transported and concentrated over South China, resulting in more precipitation over this region in early summer. The release of the latent heat from the precipitation contributes to warming the atmosphere, leading to the westward shift of the WPSH and eastward movement of the SAH. Under the control of the WPSH, South China experiences a hot and dry climate. Meanwhile, North China is at the northern edge of the WPSH, where is the convergence area of the warm wet air mass and cold air, consequently receives more precipitation. Additionally, the La Niña-like pattern difference of equatorial Pacific SST also contributes to the strong EASM in the warm period.

In this study, the spring insolation change and the response of the atmospheric system are considered to be directly responsible for the difference of the EASM precipitation between BA and YD period. The factors in low latitudes are the main focus while that in high latitudes are also important. Especially, the sea ice cover change in the high latitudes is regarded as one of the main forcing factors of climate changes due to its impact on the radiation balance and the ocean thermal circulation; e.g., the changes in Arctic sea ice extent caused by a weak AMOC are thought to be closely related to the YD formation, and the westerlies play the bridging role for connecting North Atlantic and the East Asia climate. In addition, the southward expansion of cold air is an essential element for the seasonal process of the EASM. Other aspects not considered here may relate to synoptic-scale changes in response to drastic changes over the North Atlantic region such as shifts in NAO or AO signals which could propagate along the westerly belt to East Asia and influence the climate of subtropical East Asia and tropical Pacific through the outbreak and southward invasion of the winter monsoon. Furthermore, the spring sensible heating of the Tibetan Plateau is heavily dependent on its thermal condition in

winter (Qian et al., 2002; Qiang et al., 2004; Zhou et al., 2007; Liu et al., 2014b), indicating the impact of the factors in low latitudes may indirectly and partly reflect the influences from high latitudes. Therefore, it is highly important to further study the response of the EASM evolution to the environmental changes in high latitudes and the related physical mechanism.

Additionally, we point out that the warming in BA is asymmetrical and regional, but the warming in the early Holocene is global. It implies that regional heterogeneity of warming may lead to different response of the EASM due to the complex feedback and interaction of the climate system. Therefore, further comparison of the EASM evolution under different conditions of warming are needed.

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Declarations

Conflict of interest The authors declare that they have no conflict of interest.

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