Simulation of the East Asian Summer Monsoon during the Last Millennium with the MPI Earth System Model

WENMIN MAN
LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences, and Graduate University of Chinese Academy of Sciences, Beijing, China

TIANJUN ZHOU
LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China

JOHANN H. JUNGCLAUS
Max Planck Institute for Meteorology, Hamburg, Germany

ABSTRACT

The decadal–centennial variations of East Asian summer monsoon (EASM) and the associated rainfall change during the past millennium are simulated using the earth system model developed at the Max Planck Institute for Meteorology. The model was driven by up-to-date reconstructions of external forcing including the recent low-amplitude estimates of solar variations. Analysis of the simulations indicates that the EASM is generally strong during the Medieval Warm Period (MWP; A.D. 1000–1100) and weak during the Little Ice Age (LIA; A.D. 1600–1700). The monsoon rainband exhibits a meridional tripolar pattern during both epochs. Excessive (deficient) precipitation is found over northern China (35°–42°N, 100°–120°E) but deficient (excessive) precipitation is seen along the Yangtze River valley (27°–34°N, 100°–120°E) during the MWP (LIA). Both similarities and disparities of the rainfall pattern between the model results herein and the proxy data have been compared, and reconstructions from Chinese historical documents and some geological evidence support the results. The changes of the EASM circulation including the subtropical westerly jet stream in the upper troposphere and the western Pacific subtropical high (WPSH) in the middle and lower troposphere are consistent with the meridional shift of the monsoon rain belt during both epochs. The meridional monsoon circulation changes are accompanied with anomalous southerly (northerly) winds between 20° and 50°N during the MWP (LIA). The land–sea thermal contrast change caused by the effective radiative forcing leads to the MWP and LIA monsoon changes. The “warmer land–colder ocean” anomaly pattern during the MWP favors a stronger monsoon, while the “colder land–warmer ocean” anomaly pattern during the LIA favors a weaker monsoon.

1. Introduction

The East Asian summer monsoon (EASM) is an important component in the global climate system. Its anomalous behavior leads to deficient or excessive precipitation and hence causes great economic and social losses in East Asian regions [see reviews by Wang (2006) and Zhou et al. (2009a)]. The EASM exhibits considerable variability on a wide range of time scales. Many studies have focused on the interannual or interdecadal variability of the EASM (Chang et al. 2000a,b; Wang et al. 2000; Wu et al. 2003; Yu et al. 2004; Wu et al. 2009a,b; Zhou et al. 2009b; Li et al. 2010). However, the behavior of monsoon variability on the decadal–centennial time scale during the last millennium is less examined and largely unknown.

Proxy data derived from Chinese historical documents and speleothem records have been used to reconstruct the past EASM variability, as well as the spatial patterns and temporal evolutions of precipitation over eastern China (Wang et al. 1987; Qian et al. 2003; Zheng et al. 2009).
2006; Zhang et al. 2008). A dataset from a 120-station drought/flood (D/F) index (a five-grade category index) was derived from Chinese historical documents for A.D. 1470–1979 (CMA 1981). Based on the documentary data, Wang et al. (1981) reported that the anomaly of summer rainfall over northern China (35°–42°N, 100°–120°E) was usually opposite to that over the lower-middle Yangtze River valley (27°–34°N, 100°–120°E). Based on the extended dataset of the D/F index for the period of A.D. 950–1991, it was inferred that there was more flooding in northern China during the Medieval Warm Period (MWP), while a similar condition was found along the Yangtze river valley during the Little Ice Age (LIA) (Wang et al. 1987). The flood frequency anomalies over the Yellow River valley (33°–40°N, 105°–120°E) and southern China (22°–29°N, 108°–120°E) in the warm season (May–September) during A.D. 991–1999 indicate that the flood frequency was small (1.15 decade⁻¹) over the Yellow River valley and large (2.45 decade⁻¹) over southern China during A.D. 1400–1600, which corresponds to a weakened EASM in this period. The flood frequency markedly increased over the Yellow River valley and decreased over southern China after A.D. 1650, indicating a stronger EASM (Qian et al. 2006). Zheng et al. (2006) found that the precipitation variation in eastern China exhibited dry/wet fluctuations on centennial time scales. Droughts dominated in the twelfth to fourteenth centuries, but since the middle of the seventeenth century eastern China has been more subject to flooding. However, Zhang et al. (2008) indicated that the EASM was strong during the MWP and generally wet over eastern China, whereas the LIA was characterized by a period of weak EASM and drought over all of eastern China. The Swiss record of Alpine glaciology also captures a generally strong summer Asian monsoon during the MWP and the prominent glacial advances correlate with a weakening summer monsoon during the LIA (Holzhauser et al. 2005). Since the reconstructions have to rely on relatively sparse data sources, there exist controversial issues in understanding the EASM variability and the associated rainfall patterns over eastern China.

Climate models driven by external forcing agents can provide important insights for detecting the monsoon variability on the decadal–centennial time scale. Studies of these issues will improve our understanding of the physical processes that determine the long-term monsoon variations. These kinds of simulations have been done using a wide range of climate models with different levels of complexity for the last millennium (Crowley 2000; Gerber et al. 2003; Gonzalez-Rouco et al. 2003; Goosse et al. 2005; Ammann et al. 2007; Peng et al. 2009; Jungclaus et al. 2010; Servonnat et al. 2010). However, previous studies of model–data intercomparison mainly focused on variations of surface temperatures (Stouffer et al. 2000; Zorita et al. 2005; Wagner et al. 2005; Zhang et al. 2011) and major modes of climate variation such as ENSO (Mann et al. 2005) or the North Atlantic Oscillation (Shindell et al. 2003). There has been little attempt to interpret the causes and dynamics of the decadal–centennial EASM variations using the model simulations.

Liu et al. (2011) investigated the centennial–millennial variation of the EASM precipitation over the past 1000 years through the analysis of a millennium simulation of the coupled ECHAM and the global Hamburg Ocean Primitive Equation (ECHO-G) model. The model results indicate that the centennial–millennial variation of the EASM is essentially a forced response to the external radiative forcing. The climate response of EASM to the external radiative forcing depends on latitude. However, multimodel intercomparisons are still needed to investigate whether this phenomenon is model dependent. The present study aims to examine the EASM changes during the MWP and LIA by using the millennium simulations with a comprehensive earth system model that, in contrast to previous studies, offers an (albeit small) ensemble of simulations over the last millennium. The main motivation of the study is to address the following questions: 1) What are the spatial structures of the EASM during the MWP and LIA? 2) What are the forced responses of the East Asian summer rainfall during the two epochs? How about the consistency between the simulation and proxy data? 3) What is the dominant reason for the centennial EASM changes? The remainder of the paper is organized as follows. Section 2 provides a description of the model and the experimental design, as well as the details of external forcings used in the simulations. Section 3 presents the results. The conclusions are given in Section 4 along with a discussion.

2. Model and data description

a. Model description

The present study is based on the millennium experiments using the Max Planck Institute for Meteorology Earth System Model (MPI-ESM) (Jungclaus et al. 2010). The model consists of the atmospheric general circulation model ECHAM5 (Roeckner et al. 2003) and the Max Planck Institute Ocean Model (MPI-OM; Marsland et al. 2003). Modules for land vegetation [Jena Scheme for Biosphere–Atmosphere Coupling in Hamburg (JSBACH); Raddatz et al. 2007] and ocean biogeochemistry [Hamburg Model of the Ocean Carbon Cycle (HAMOCC); Wetzel et al. 2006] enable the interactive simulation of the carbon cycle. ECHAM5 is run at T31 resolution (~3.75°)
with 19 vertical levels, resolving the atmosphere up to 10 hPa. MPI-OM applies a conformal mapping grid with a horizontal resolution ranging from 22 to 350 km. The ocean model includes a Hibler-type dynamic–thermodynamic sea ice model with viscous plastic rheology (Hibler 1979). Ocean and atmosphere are coupled daily without flux corrections using the Ocean Atmosphere Sea Ice Soil version 3 (OASIS3) coupler (Valcke et al. 2003).

b. Experimental design and forcing data

The experimental strategy is described as follows. After a multicentury spinup phase in which the carbon cycle was brought into equilibrium, a 3000-yr unforced control experiment was performed under A.D. 800 orbital conditions and preindustrial greenhouse gas concentrations. Starting from different ocean initial conditions, a five-member ensemble (E1) with the standard external forcing spanning A.D. 800–2005 was conducted using the earth system model.

The total solar irradiance (TSI) forcing used as the standard forcing exhibits an increase of 0.1% (~1.3 W m$^{-2}$) from the Maunder Minimum to today (Viera et al. 2011), which is in agreement with other recent evaluations, although other reconstructions with higher long-term variations also exist [see the discussion in Schmidt et al. (2011)]. The volcanic forcing is calculated online in the model using time series of aerosol optical depth and of the effective radius (Crowley et al. 2008). Anthropogenic land cover change is considered by applying the reconstruction of global agricultural areas and land cover (Pongratz et al. 2008). While the CO$_2$ concentration is calculated interactively within the model, the concentrations of the other two major greenhouse gases, methane (CH$_4$) and nitrous oxide (N$_2$O), are prescribed (MacFarling Meure et al. 2006). Some other potentially important forcings such as the orbital forcing and anthropogenic tropospheric sulfate aerosols are also included in the ensemble experiments. The orbit forcing has little effect on the magnitude of the seasonal cycle [see Jungclaus et al. (2010) for details].

Effective radiative forcings (Fig. 1) are calculated offline with the ECHAM5 isolated radiative transfer code following the Wetherald and Manabe (1998) approach for calculating radiative feedbacks. The anomalous total radiative forcing, which represents the sum of the solar forcing and the radiative effects of volcanic aerosols, land cover change, and the CO$_2$ concentration, follows a high value during the MWP and a low value during the LIA in the simulations (Fig. 1).

c. Data

The data used for validation of the model performance under the present-day climate include the following:

1) The precipitation dataset compiled by Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) for the period of 1979–2005 on a 2.5° × 2.5° grid (Xie and Arkin 1995); and
2) The National Centers for Environmental Prediction (NCEP) reanalysis 2 data (NCEP2) for 1979–2005 on a 2.5° × 2.5° grid (Kanamitsu et al. 2002).

3. Results

We first present the summer [June–August (JJA)] rainfall distribution and seasonal cycle of precipitation on an observational basis for the model validation. Then we focus on the large-scale monsoonal circulation and precipitation changes during the MWP and LIA. Finally, we try to attribute the underlying causes of the EASM variations from the perspective of land–sea thermal contrast. In the following discussions, all the anomalies are calculated relative to the climate mean of the whole millennium.
Faithful analysis of the EASM response to external forcing should be based on rigorous verification of the model performance (Liu et al. 2011). Comparisons have demonstrated that the simulated Northern Hemisphere temperature evolution and regional temperature anomalies in China agree well with the reconstructions (Jungclaus et al. 2010; Zhang et al. 2011). We focus on the mean state of summer rainfall as well as the seasonal march of the monsoon rain belt to evaluate the performance of the model in monsoon simulation in this study. Variations of the EASM are usually described using summer rainfall and low-level (850 hPa) winds (Zhou and Yu 2005; Yu and Zhou 2007; Chen et al. 2010). Figures 2a and 2b compare the simulated summer rainfall and 850-hPa winds for the average of 1979–2005 with the observations. The observed precipitation data are derived from CMAP, while the circulation data are from NCEP2. The summer 850-hPa winds feature strong southwesterlies from the Indian monsoon and southeasterlies from the western Pacific, as well as the cross-equator flow around 105°–120°E (Fig. 2a). The observed summer precipitation generally decreases northwestward from the Southeast Asian marginal seas toward the arid central continental Asia, and there is a monsoon rainband extending from the East Asian marginal continents to Japan (Fig. 2a). The model captures the main features of the observed summer precipitation realistically except that it underestimates the monsoon rainband extending from eastern China to Japan (Fig. 2b). The monsoon wind penetrates into northern China in the model, resulting in a northward bias of the monsoon rainband. An artificial rainfall center located to the eastern periphery of the Tibetan Plateau is also evident in the simulation, as in many atmospheric general circulation models (AGCMs) (Yu et al. 2000; Zhou and Li 2002; Chen et al. 2010). The pattern correlation coefficient of the summer precipitation between the observation and model simulation is 0.81, and the root-mean-square difference between the observation and the simulation is 2.24 mm day\(^{-1}\).

East Asian precipitation exhibits a robust seasonal cycle associated with monsoon development (Zhou et al. 2009a). The seasonal cycles of extratropical (36°–50°N,
100°–120°E) and subtropical (21°–35°N, 100°–120°E) precipitation are shown in Figs. 2c and 2d. In extratropical East Asia (EA; Fig. 2c), the seasonal cycle is well simulated with a peak rainy month in July and a minimum in January, but the simulated spring rainfall is stronger than that in the observation. In subtropical East Asia (Fig. 2d), the simulation resembles the observation in the June peak but the strength of precipitation prior to June is overestimated. Overall, the model performs well in the simulation of both the seasonal cycle and precipitation amounts in different latitudinal regions of East Asia.

b. Response of EASM circulation and precipitation during the MWP and LIA

Since the focus of the study is the dynamic structure and physical processes of EASM variations on the decadal–centennial time scale, we examine the features during the MWP (A.D. 1000–1100) and LIA (A.D. 1600–1700), which represent two typical climatic epochs of the last millennium in China. We focus on the ensemble mean of five realizations in the following analysis.

Anomalies of JJA mean 850-hPa winds and precipitation during the MWP and LIA are shown in Fig. 3. The MWP (LIA) is characterized by the strengthening (weakening) of the 850-hPa southwesterly winds, indicating a generally stronger (weaker) EASM. This result is consistent with the reconstruction from a stalagmite record in the Wanxiang Cave, China, which indicates a strong (weak) EASM during the MWP (LIA) (Zhang et al. 2008).

The anomalies of monsoon rainfall exhibit a meridional tripolar pattern during both epochs. Excessive (deficient) precipitation is evident over northern China (35°–42°N, 100°–120°E) whereas deficient (excessive) precipitation is seen along the Yangtze River valley (27°–34°N, 100°–120°E) during the MWP (LIA). The rainfall patterns are consistent with the reconstructions from Chinese historical documents (Wang et al. 1987; Qian et al. 2003), which suggest that changes of rainfall along the Yangtze River valley are generally out of phase from those over northern China, but different from the reconstruction of the Wanxiang record (Zhang et al. 2008), with wet (dry) conditions over all of eastern China during the MWP (LIA). The reason for this discrepancy deserves further investigations. Some geological evidence also supports more precipitation over northern China during the MWP (Ren and Zhang 1996; Wu and Lu 2005; Cao et al. 2004). For the rainfall along the Yangtze River valley, the reconstructed 3000-yr precipitation curve from the Longgan Lake (29°50′–30°05′N, 115°55′–116°20′E) shows that a drier climate locally occurred in most of the MWP (Tong et al. 1997). Our result is also in accordance with the modern definitions of the EASM in that when southerly monsoon penetrates deeply into northern China, extratropical EA has plentiful rainfall, but the subtropical rainfall (known as mei-yu in China, Baiyu in Japan, and Changma in Korea) tends to be suppressed (Ding 1992). Note this definition is different to that from the stalagmite proxy data (i.e., a strong EASM often means an abundant mei-yu; Wang et al. 2005). Our model result agrees with the observed structure on interannual time scale. We also calculate the ensemble spread of precipitation during the two epochs, and define the signal-to-noise ratio as the absolute value of the ensemble mean value dividing the ensemble spread (Zhou and Yu 2006). The signal-to-noise ratio is greater than 1.0 over most parts of the EA region (not shown), indicating that the magnitude of the forced signal is larger than the model spread, which support the robustness of the responses in our simulation.
c. The horizontal circulation of EASM during the MWP and LIA

The water vapor transport is crucial to monsoon rainfall. Since the vertically integrated water vapor transport is dominated by the lower troposphere (Zhou and Yu 2005), an analysis of water vapor transport at 850 hPa is reasonable. There are three main branches of climatological 850-hPa water vapor transport to EA (Fig. 4a): a strong transport by the southwesterlies from the Indian monsoon, a moderate transport by the Southeast Asian monsoon from the western Pacific, and a weak transport of cross-equator flow straddling 105°–150°E. Anomalies of JJA mean 850-hPa water vapor transport during the MWP and LIA are shown in Figs. 4b and 4c. The anomalous water vapor transport by both the southwesterlies from the Indian monsoon and the southeasterlies from the western Pacific are positive during the MWP and, in addition to the anomalous southerly winds over EA, the northward transport of tropical water vapor to northern China has enhanced, but the water vapor convergence over the Yangtze River valley has decreased (Fig. 4b). This leads to excessive rainfall in northern China and deficient rainfall in central China along the Yangtze River valley. The northward moisture transport has reduced during the LIA and there is more water vapor convergence over the Yangtze River valley (Fig. 4c), resulting in excessive rainfall in central China but deficient rainfall in northern China.

To quantify the water vapor transports, estimations of the atmospheric water budget across the four boundaries of northern China and the Yangtze River valley have been calculated according to Li et al. (2009). During the MWP, the 850-hPa net influx in northern China was 6.25 \times 10^6 \text{ kg s}^{-1}, which was much larger than that over the Yangtze River valley (2.12 \times 10^6 \text{ kg s}^{-1}), resulting in excessive rainfall in northern China but deficient rainfall in central China along the Yangtze River valley. During the LIA, the 850-hPa net influx in northern China (1.21 \times 10^6 \text{ kg s}^{-1}) was much smaller than that over the Yangtze River valley (6.05 \times 10^6 \text{ kg s}^{-1}). This leads to deficient rainfall in northern China and excessive rainfall along the Yangtze River valley.

The water vapor transport is closely linked to the monsoon circulation change. Previous studies have found that the EASM is greatly controlled by the western Pacific subtropical high (WPSH) in the middle and lower troposphere and the subtropical westerly jet stream in the upper troposphere (Tao and Chen 1987). The position, shape, and strength of the WPSH dominate the large-scale quasi-stationary frontal and associated rainband in EA (Tao and Chen 1987; Ding 1994; Zhou and Yu 2005). The WPSH is conventionally measured by the geopotential height at 500 hPa (Tao and Chen 1987; Ding 1994; Zhou et al. 2009c). In Fig. 5, we present the changes of geopotential height of 500 hPa for both the MWP and LIA. It is clear that the ridge of the geopotential isolines shift northward (southward) during the MWP (LIA) compared with the climate mean position (Figs. 5a,b). The northward (southward) shift of the WPSH ridge during the MWP (LIA) leads to stronger (weaker) southerlies penetrating into northern China, resulting in deficient (excessive) precipitation along the Yangtze River Valley but excessive (deficient) precipitation over northern China.

At the upper levels (200 hPa), the subtropical westerly jet is an important part of the Tibetan high, the position and strength of which is closely related to the EASM rainfall (Zhang et al. 2006). The most outstanding features of the climatological EASM at 200 hPa
are the westerly jet stream centered along 40°N and the tropical easterly jet to the south of 25°N (Fig. 6a). Anomalies of JJA mean 200-hPa zonal winds during the MWP and LIA are shown in Figs. 6b and 6c. The subtropical westerly jet exhibits apparent meridional shifts during the two epochs. There is a significantly weakened (intensified) westerly south to the jet axis and an intensified (weakened) westerly north to the jet axis during the MWP (LIA). This corresponds to the northward (southward) shift of the monsoon rain belt and is generally accompanied by excessive (deficient) rainfall over northern China but deficient (excessive) rainfall along the Yangtze River valley (Lau et al. 1988; Liang and Wang 1998; Li et al. 2004; Zhou and Yu 2005).

d. Meridional monsoon circulation changes during the MWP and LIA

The Hadley cell is a thermally driven meridional circulation, which is characterized by a rising motion in the tropics and a descending motion in the subtropics. The normal Hadley cell in the East Asian monsoon region is replaced by a meridional circulation of the opposite sense, which is often referred to as the monsoonal meridional cell (Chen et al. 1964; Ye and Yang 1979) and has been used as observational metric in model evaluations (Zhou and Li 2002; Chen et al. 2010). To examine the meridional structure of the EASM during the MWP and LIA, Fig. 7 shows a meridional–vertical cross section of JJA mean winds along 105°–122°E. The climatological map exhibits a strong upward motion over Southeast Asia (~5°–30°N) and a weak ascent north of about 40°N (Fig. 7a). The anomalous meridional monsoon

Fig. 5. Spatial distributions of JJA mean geopotential height at 500 hPa for (a) the MWP and (b) the LIA (both shown in color). The results of the climatological millennial mean are shown in long-dashed black. Contour interval is 10 gpm.

Fig. 6. JJA mean 200-hPa zonal wind (m s⁻¹) for (a) the millennial mean, (b) the MWP, and (c) the LIA. The anomalies of the MWP and LIA are calculated relative to the millennial mean value. The shading denotes regions that are statistically significant at the 5% level by using a Student’s t test.
circulation during the MWP shows upward motion over the extratropical EA and downward motion over the subtropical EA. This corresponds to anomalous low-level southerlies between 20° and 50°N (Fig. 7b). Anomalous descending and ascending motion respectively dominate the northern land and southern ocean areas during the LIA, with anomalous northerly winds between 20° and 50°N (Fig. 7c). The meridional circulation changes in the monsoon region are consistent with an enhanced (weakened) EASM during the MWP (LIA).

**e. The driving mechanisms of EASM change:**

**Land–sea thermal contrast**

The occurrence of the EASM variability is a consequence of the atmospheric response to the diabatic heating between the ocean and the land (Li and Yanai 1996). The spatial structure of the monsoon response can be understood in terms of the effects of land–sea thermal contrast. Studies of the monsoon variability from this perspective can help us understand the physical and dynamical processes that determine the long-term monsoon variations.

The land–sea thermal contrast is attributable to external forcings. We term the effective radiative forcing from the external drivers as the sum of the effective solar irradiance, the radiative effects of volcanic aerosols, the land-cover changes, and greenhouse-gas forcing. All of the individual effective radiative forcings are calculated offline with the ECHAM5 isolated radiative transfer code. The effective solar irradiance has been multiplied by planetary albedo (we simply use a constant of 0.7) and divided by 4. The effective radiative forcing anomaly between the MWP and LIA differ by 0.23 W m\(^{-2}\) in the simulation. The calculation of the individual component differs by 0.03 W m\(^{-2}\) for the solar forcing and 0.19 W m\(^{-2}\) for the volcanic forcing over the 100-yr periods for the MWP and LIA. The result indicates that volcanic forcing has a major contribution for the total effective radiative forcing difference between the MWP and LIA. The anomaly of the individual effective radiative forcing is −0.01 W m\(^{-2}\) for solar forcing and −0.26 W m\(^{-2}\) for volcanic forcing over the 100-yr periods for the LIA compared with the long-term mean. The calculation of the mean radiative forcing for the anomalies of the MWP and LIA are calculated relative to the millennial mean value. The gray shading denotes regions with vertical components that are statistically significant at the 5% level by using a Student’s \(t\) test.

![Fig. 7. Latitude–height cross section of JJA mean meridional circulation averaged over 105°–122°E (units of the vertical and meridional velocity are \(-10^{-2}\) hPa s\(^{-1}\) and m s\(^{-1}\), respectively) for (a) the millennial mean, (b) the MWP, and (c) the LIA. The...](image-url)
LIA further indicates that the LIA is dominated by the volcanic forcing. When the effective radiative forcing increases during the MWP because of the different thermal capacity of the land and ocean, the temperature increases over the East Asian continent, especially in the midlatitude, more rapidly than over the adjacent ocean, which produces the land–sea thermal contrast during the period.

The tropospheric mean temperature is a reasonable indicator of thermal contrast change (Zhou and Zou 2010). The tropospheric mean (200–500-hPa average) temperature anomalies during the MWP and LIA are shown in Fig. 8. Warm temperature anomalies prevail over EA with a central magnitude of 0.2°C during the MWP, while cool anomalies are seen in the tropical western Pacific and extratropical North Pacific with a central value of −0.25°C (Fig. 8a). There exist cold anomalies with an amplitude up to −0.3°C over EA and warming anomalies with a central magnitude of 0.25°C in the tropical western Pacific and extratropical North Pacific during the LIA (Fig. 8b). The magnitude of temperature changes during the LIA is stronger than the MWP. The position of the cooling center over EA exhibits a northward displacement during the LIA. Since the mean state of summertime tropospheric mean temperature features a “warm land–cold ocean” condition, the “warmer land–colder ocean” anomaly pattern during the MWP favors a strong EASM circulation, while the “colder land–warmer ocean” anomaly pattern during the LIA favors a weak EASM circulation. Thus the land–sea thermal contrast change is the fundamental driver of EASM changes.

The EASM change is caused by both zonal thermal contrast between EA and the North Pacific and meridional thermal contrast between EA and the tropical western Pacific (Fig. 8). To have a clear picture of both zonal and meridional land–sea thermal contrast change, the corresponding structures at vertical cross sections are shown in Figs. 9 and 10. The zonal land–sea thermal contrast along 30°–45°N depicted by the height–longitude cross section (Fig. 9) exhibits a signature of warmer land–colder ocean during the MWP, with warmer anomalies of 0.2°C over the Eurasian continent extending from 60° to 120°E. The cooler anomalies with a central magnitude of −0.2°C are seen over the ocean area in the middle-upper troposphere, extending from 120°E to 150°W. Temperature anomalies of almost reversed sign are evident during the LIA, with negative anomalies over land extending from 75° to 120°E and positive anomalies with an amplitude up to 0.3°C over an ocean area extending from

**FIG. 8.** JJA mean upper-tropospheric (500–200 hPa) temperature anomalies (°C) for (a) the MWP and (b) the LIA. The anomalies are calculated relative to the millennial mean value. The black stippled regions denote areas that are statistically significant at the 5% level by using a Student’s *t* test.

**FIG. 9.** Longitude–height cross section of JJA temperature averaged over 30°–45°N (°C) for (a) the MWP and (b) the LIA. The anomalies are calculated relative to the millennial mean value. The black stippled regions denote areas that are statistically significant at the 5% level by using a Student’s *t* test.
The simulated colder land is weaker in magnitude (\(0.1^\circ C\)) and narrower in zonal extent, exhibiting the maximum magnitude in the middle-lower troposphere below 500 hPa. However, the warmer anomalies over ocean area show a deep vertical structure and penetrate throughout the troposphere.

The height–latitude cross section measuring the meridional land–sea thermal contrast also exhibits a warmer land–colder ocean structure during the MWP (Fig. 10a). The “warmer land” is evident with its maximum warming of \(0.3^\circ C\) around 300–500 hPa extending from 30° to 60°N. The temperature over the ocean is also warmer during the MWP, but the magnitude is weaker than that over the land, so the land–sea thermal contrast still increases. The colder land–warmer ocean structure is evident in the troposphere during the LIA (Fig. 10b), with colder anomalies of \(-0.2^\circ C\) over land extending from approximately 30°–60°N and warmer anomalies of \(0.2^\circ C\) over ocean extending from 0° to 30°N.

Following the changes of temperature gradients, anomalous descending and ascending motion dominate the northern land and southern ocean areas, respectively, with anomalous low-level northerly winds between 20° to 50°N.

The strongest cooling center is around the 300-hPa level with a magnitude of \(-0.2^\circ C\). The cooling center results in a southward shift of high-level subtropical westerly jet while enhancing the anomalous low-level northerlies, similar to what happened in the later twentieth century associated with a weakened EASM (Yu et al. 2004; Yu and Zhou 2007). In addition, the change of meridional temperature gradient is also consistent with the changes of subtropical westerly jet shown in Fig. 6, a poleward increase (decrease) of tropospheric temperature is followed by a weakened (intensified) westerly, based on the principle of thermal wind balance (Zhang et al. 2006).

In summary, the spatial structure of land–sea thermal contrast and the associated circulation changes reasonably explains an enhanced (weakened) EASM during the MWP (LIA), which is further confirmed by the differences between MWP and LIA (Fig. 11), namely that the stronger EASM during MWP relative to LIA is dominated by the enhanced land–sea thermal contrast. The spatial pattern of MWP–LIA tropospheric mean temperature differences reveals warmer anomalies with an amplitude up to 0.45°C over EA and colder anomalies with a central magnitude of \(-0.5^\circ C\) in the tropical western Pacific and extratropical North Pacific. Following the land–sea thermal contrast change, the southerlies penetrate northward into higher latitudes in the MWP than in the LIA.

4. Summary and discussion

a. Summary

The EASM changes and the corresponding rainfall patterns over EA during the MWP and LIA are analyzed by using the output of the MPI Earth System Model. The association between the EASM and land–sea thermal contrast changes is studied. The main results are summarized below:

1) The EASM during the MWP is stronger than that during the LIA, as the proxy data indicated. Following the intensified (weakened) EASM during the MWP (LIA), the summer precipitation over eastern China exhibits a meridional tripolar pattern. Excessive (deficient) rainfall is found over northern China but deficient (excessive) rainfall along the Yangtze River valley during the MWP (LIA). Both similarities and disparities between our model results and the available estimates from the proxy data have been compared: reconstructions from Chinese historical documents and some geological evidence support our results, but reconstructions from the Wangxiang record show disparities with the model results.

Fig. 10. Latitude–height cross section of JJA temperature averaged over 105°–122°E (°C) for (a) the MWP and (b) the LIA. The anomalies are calculated relative to the millennial mean value. The black stippled regions denote areas that are statistically significant at the 5% level by using a Student’s \(t\) test.

120°E to 150°W. The simulated colder land is weaker in magnitude (\(-0.1^\circ C\)) and narrower in zonal extent, exhibiting the maximum magnitude in the middle-lower troposphere below 500 hPa. However, the warmer anomalies over ocean area show a deep vertical structure and penetrate throughout the troposphere.

The height–latitude cross section measuring the meridional land–sea thermal contrast also exhibits a warmer land–colder ocean structure during the MWP (Fig. 10a). The “warmer land” is evident with its maximum warming of 0.3°C around 300–500 hPa extending from 30° to 60°N. The temperature over the ocean is also warmer during the MWP, but the magnitude is weaker than that over the land, so the land–sea thermal contrast still increases. The colder land–warmer ocean structure is evident in the troposphere during the LIA (Fig. 10b), with colder anomalies of \(-0.2^\circ C\) over land extending from approximately 30°–60°N and warmer anomalies of \(0.2^\circ C\) over ocean extending from 0° to 30°N.

Following the changes of temperature gradients, anomalous descending and ascending motion dominate the northern land and southern ocean areas, respectively, with anomalous low-level northerly winds between 20° to 50°N.
2) The northward water vapor transport has enhanced (reduced) during the MWP (LIA), which leads to less (more) moisture convergence and thus deficient (excessive) rainfall along the Yangtze River valley but excessive (deficient) rainfall in northern China. Both the changes of the subtropical westerly jet stream in the upper troposphere and the WPSH in the middle and lower troposphere are consistent with the meridional shift of the monsoon rain belt during both epochs. A stronger (weaker) WPSH along with a northward (southward) shift of the subtropical westerly jet stream is evident in the warm (cold) period.

3) The meridional monsoon circulation changes show anomalous ascending (descending) and descending (ascending) motion respectively over the northern land and southern ocean areas during the MWP (LIA). An anomalous southerly (northerly) wind is seen between 20° and 50°N, which corresponds to an enhanced (weakened) summer monsoon.

4) The land–sea thermal contrast changes caused by the effective radiative forcing lead to the MWP and LIA monsoon changes. The EASM is dominated by both the zonal thermal contrast between EA and the North Pacific and the meridional thermal contrast between EA and the tropical western Pacific. The “warmer land–colder ocean” anomaly pattern during the MWP favors a strong EASM, while the “colder land–warmer ocean” anomaly pattern during the LIA favors a weak EASM.

b. Discussion

Based on the output of MPI Earth System Model millennial climate simulations, our analysis shows that the EASM during the MWP is stronger than that during the LIA. There is excessive (deficient) rainfall in northern China but deficient (excessive) rainfall along the Yangtze River valley during the MWP (LIA). We compared our results with the available estimates derived from the proxy data. Reconstructions from Chinese historical documents and some geological evidence support our results, but reconstructions from the Wanxiang record show disparities with the model results. The reason deserves further investigation.

The model output analyzed in Liu et al. (2011) was obtained from the millennium integrations of the coupled ECHO-G model, which consists of the spectral atmospheric model ECHAM4 and the global ocean circulation model HOPE-G. The simulation was forced by three external forcing factors: solar variability (Crowley 2000), greenhouse gas concentrations in the atmosphere including CO₂ (Etheridge et al. 1996) and CH₄ (Blunier et al. 1995), and an estimated radiative effect of volcanic aerosols (Robock and Free 1996). Note that the level of solar irradiance used to drive the model in Liu et al. (2011) exhibits larger change in comparison to the “state of the art” estimates for solar variability applied in this study. Both our result and the work by Liu et al. (2011) indicate that centennial variation of the EASM is essentially a forced response to the external radiative forcing. Besides the similarities, however, there are also disparities between the two simulations. The results from our study indicate that the summer precipitation over eastern China exhibits a meridional tripolar pattern, which is characterized by an out-of-phase relationship between the subtropical and extratropical rainfall. This feature is consistent with the observed structure on the interannual time scale. The subtropical and extratropical rainfall increases simultaneously in Liu et al. (2011). This feature implies that the forced mode is characterized by an in-phase relationship between the subtropical and extratropical rainfall. The main discrepancy in the rainfall between our study and the work by Liu et al. (2011) is in the Yangtze River valley whereas the extratropical (corresponding to Northern China) response is consistent. Since the warming in high latitudes is much larger than that over the tropical regions during the MWP, both of the models...
show enhanced southerly winds across the subtropical and extratropical monsoon regions, indicating that the responses of the circulations within the two models are similar. However, the precipitation could differ significantly even though the atmospheric circulations are the same between different models. We suggest that the disparities of the precipitation in the two simulations could be model dependent; thus, multimodel intercomparisons are needed in future studies. We should note that the subtropical region (21°–35°N, 100°–120°E) in Liu et al. (2011) covers southern China and the Yangtze River valley, and excessive (deficient) rainfall is found over southern China but deficient (excessive) rainfall along the Yangtze River valley during the MWP (LIA) in our simulation. If we calculate the regional-mean precipitation value for the subtropical region defined by Liu et al. (2011), we also find that subtropical precipitation was strong during MWP and weak during LIA.

We use the model output from the full-forcing experiments in this study, so it is difficult to assess responses from one specific forcing component. However, examination of the specific response from one individual forcing is quite important; for example, the land-use forcing could be an important aspect for regional climate during the last millennium. By using an atmospheric general circulation model, Takata et al. (2009) suggested that the land cover/use change between 1700 and 1850 could result in the weakening of the Asian summer monsoon through changes in the energy and water balance at Earth’s surface. Thus, the effect of land-use forcing on the regional climate deserves further study by experiments in which just one forcing component is applied.

A map with the change in land use between the MWP and LIA is further provided in order to discuss possible local reinforcement of the land–sea contrast due to land use between the two periods (Fig. 12). Large areas of cropland were deduced for China during the MWP (LIA) in our simulation. If we calculate the regional-mean precipitation value for the subtropical region defined by Liu et al. (2011), we also find that subtropical precipitation was strong during MWP and weak during LIA.

We use the model output from the full-forcing experiments in this study, so it is difficult to assess responses from one specific forcing component. However, examination of the specific response from one individual forcing is quite important; for example, the land-use forcing could be an important aspect for regional climate during the last millennium. By using an atmospheric general circulation model, Takata et al. (2009) suggested that the land cover/use change between 1700 and 1850 could result in the weakening of the Asian summer monsoon through changes in the energy and water balance at Earth’s surface. Thus, the effect of land-use forcing on the regional climate deserves further study by experiments in which just one forcing component is applied.

A map with the change in land use between the MWP and LIA is further provided in order to discuss possible local reinforcement of the land–sea contrast due to land use between the two periods (Fig. 12). Large areas of cropland were deduced for China during the MWP (LIA). The spread of crops further developed during the LIA compared with the MWP (Fig. 12b). The distribution of pasture was quite different from that of cropland (Figs. 12c,d). The pasture lands during both periods were mainly located in Mongolia and Tibet, where herding was the traditional form of agriculture. Many parts of eastern China showed little pasture area during those two
periods. The total areas of cropland and pasture were larger during the MWP than that during the LIA over this region. The land-use change would favor a land–sea thermal contrast change between the two periods.

We state the major features of symmetries between the patterns during the MWP and LIA, and it is interesting to see that it is not entirely symmetrical. The anomaly of the effective radiative forcing is $-0.03 \text{ W m}^{-2}$ for the MWP and $-0.26 \text{ W m}^{-2}$ for the LIA compared to the long-term mean value, which is not symmetrical in amplitude between MWP and LIA. The asymmetry of the effective radiative forcing between MWP and LIA is consistent with the asymmetrical responses between the two typical periods. The possible reasons for the asymmetry in the patterns are both important and interesting. The asymmetry between the MWP and LIA in the rainfall pattern displays as significant negative–positive–negative anomaly pattern extending from northwestern China to the East Asian marginal continents. Corresponding to the rainfall pattern, a wave train structure is evident with negative and positive anomalies by turns in the geopotential height at 200 hPa (not shown). The wave structure of the 200-hPa geopotential height is consistent with the asymmetry between the MWP and LIA in the rainfall pattern. The asymmetry of the patterns and possible reasons for the asymmetry still deserve further diagnosis in our future study.

Additionally, our study shows evidence that the changes of land–sea thermal contrast associated with the effective radiative forcing dominate the EASM response. Previous studies suggest that the direct radiative effect of solar forcing variations on the monsoon change is relatively weak and that dynamical responses may be more important (Fan et al. 2009). Major climate modes during different epochs, such as ENSO and the Pacific decadal oscillation (PDO), may also contribute to the changes of EASM and deserves further study. In addition, the current diagnosis is based on the MPI model driven by a solar irradiance reconstruction of weaker variability, and thus the forcing of effective solar irradiance to EASM variation on decadal–centennial time scales may be underestimated. To account for uncertainty in the solar forcing, another set of three-member ensemble simulations (E2) with an alternative solar irradiance reconstruction of stronger variability has been done in MPI and the results will be used in our future diagnosis.

Furthermore, the EASM changes during the MWP and LIA are primarily dominated by the natural variability, such as solar and volcanic forcing variability during the last millennium. However, the future monsoon variations could be affected by human activities, such as anthropogenic forcings. Further, the EASM was stronger during the MWP, whereas the monsoon has weakened during the latter half of the twentieth century when the warming was rapid. Further study is necessary to understand the reasons behind the EASM changes under two similar climate backgrounds. Thus, it is not appropriate to suggest that the monsoon variability during the last millennium is a possible analog for future monsoon changes.

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