Global Monsoon Responses to Decadal Sea Surface Temperature Variations during the Twentieth Century: Evidence from AGCM Simulations

JIE JIANG

State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics, Institute of Atmospheric Physics, Chinese Academy of Sciences, and University of the Chinese Academy of Sciences, Beijing, China

TIANJUN ZHOU

State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics, Institute of Atmospheric Physics, Chinese Academy of Sciences (CAS), CAS Center for Excellence in Tibetan Plateau Earth Sciences, Chinese Academy of Sciences, and University of the Chinese Academy of Sciences, Beijing, China

(Manuscript received 27 December 2018, in final form 30 May 2019)

ABSTRACT

Multidecadal variations in the global land monsoon were observed during the twentieth century, with an overall increasing trend from 1901 to 1955 that was followed by a decreasing trend up to 1990, but the mechanisms governing the above changes remain inconclusive. Based on the outputs of two atmospheric general circulation models (AGCMs) forced by historical sea surface temperature (SST) covering the twentieth century, supplemented with AGCM simulations forced by idealized SST anomalies representing different conditions of the North Atlantic and tropical Pacific, evidence shows that the observed changes can be partly reproduced, particularly over the Northern Hemisphere summer monsoon (NHSM) domain, demonstrating the modulation of decadal SST changes on the long-term variations in monsoon precipitation. Moisture budget analysis is performed to understand the interdecadal changes in monsoon precipitation, and the dynamic term associated with atmospheric circulation changes is found to be prominent, while the contribution of the thermodynamic term associated with humidity changes can lead to coincident wetting over the NHSM domain. The increase (decrease) in NHSM land precipitation during 1901–55 (1956–90) is associated with the strengthening (weakening) of NHSM circulation and Walker circulation. The multidecadal scale changes in atmospheric circulation are driven by SST anomalies over the North Atlantic and the Pacific. A warmer North Atlantic together with a colder eastern tropical Pacific and a warmer western subtropical Pacific can lead to a strengthened meridional gradient in mid-to-upper-tropospheric thickness and strengthened trade winds, which transport more water vapor into monsoon regions, leading to an increase in monsoon precipitation.

1. Introduction

The global monsoon system, a global-scale atmospheric overturning circulation, represents the response of the coupled atmosphere–land–ocean system to the annual cycle of solar radiation (Trenberth et al. 2000). The global monsoon domain, identified by the annual range of precipitation, includes seven individual monsoon domains: North America, South America, northern

Africa, southern Africa, South Asia, East Asia, and Australia (Wang and Ding 2008; Wang et al. 2017). Variations in the global monsoon precipitation can affect more than two-thirds of the world's population (Wang and Ding 2006), and the changes in monsoon precipitation have enormous impacts on social and economic development over these regions.

Thermal contrast between different hemispheres caused by the annual variations in solar radiation is the basic driver of the global monsoon system. In addition, complicated topography and land–sea distribution can influence regional monsoons. Thus, the monsoons around different parts of the world have both consistent variations and regional uniqueness (Trenberth et al. 2000; Wang et al. 2017). Based on different measures of

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-18-0890.s1>.

Corresponding author: Dr. Tianjun Zhou, zhoutj@lasg.iap.ac.cn

monsoon strengths, many studies have investigated the decadal and multidecadal variations in monsoon precipitation over individual monsoon domains, most of which focused on the second half of the twentieth century due to sparse reliable datasets before the 1950s (Kripalani et al. 2003; Hu et al. 2003; Hoerling et al. 2006; Meehl and Arblaster 2011; Arias et al. 2012; Grimm and Saboia 2015; among many others). Beyond regional scales, the variations in the global monsoon have also been documented in recent years. For example, examinations of the changes in global monsoon precipitation over land found an overall weakening for the period of 1948–2003, which was mainly due to the weakening of the summer monsoon precipitation in the Northern Hemisphere land areas (Wang and Ding 2006; Zhou et al. 2008a; Zhang and Zhou 2011). The spatial pattern of precipitation trends for the period of 1930–2004 also shows a drying trend over monsoon domains, especially for the northern African and South Asian monsoon domains (Kumar et al. 2013). However, both global monsoon precipitation intensity and area significantly increased after 1979 in association with the strengthening of monsoon circulation (Hsu et al. 2011; Wang et al. 2012; Lin et al. 2014).

Why is the monsoon precipitation exhibiting such kinds of interdecadal variations? At the regional scale, numerous studies have investigated the linkage between regional monsoon precipitation and the major modes of sea surface temperature (SST) anomalies. For instance, the local Indian Ocean SST (Kucharski et al. 2006), the interdecadal variations in El Niño–Southern Oscillation (ENSO; Krishnamurthy and Goswami 2000), the Pacific decadal oscillation (PDO; Meehl and Hu 2006; Krishnamurthy and Krishnamurthy 2017), and the Atlantic multidecadal oscillation (AMO; Goswami et al. 2006; Kucharski et al. 2009; Krishnamurthy and Krishnamurthy 2016) are all important factors for the interdecadal variations in the Indian summer monsoon. The monsoon precipitation over the Sahelian region is strongly correlated with the AMO (Zhang and Delworth 2006) and the interdecadal Pacific oscillation (IPO; Mohino et al. 2011). Observational evidence also shows the relation between Asian and African monsoons at an interdecadal scale, which can be modulated by the AMO (Liu and Chiang 2012; Li et al. 2017) and the SST gradient between equatorial and extratropical regions (Feudale and Kucharski 2013). The North American monsoon is linked to the AMO (Sutton and Hodson 2005), and the Australian monsoon is linked to the IPO (Power et al. 1999).

The interdecadal variations in global monsoon precipitation have also been a research focus. Based on an ensemble of simulations with an atmospheric general circulation model (AGCM) driven by historical SSTs, it

was found that the significant weakening in global land precipitation during the period of 1949–2001 was mainly caused by the warming trend over the central–eastern Pacific and the western tropical Indian Ocean (Zhou et al. 2008b). In addition, the Northern Hemisphere summer monsoon (NHSM) circulation intensity index is highly correlated with the mega-ENSO index (Wang et al. 2013). Subtropical highs over the Northern and Southern Hemisphere Pacific, trade winds, and the Walker circulation strengthen with a warmer western Pacific and a colder eastern Pacific, leading to more precipitation in monsoon domains (Wang et al. 2017; Wang et al. 2018). The AMO is another factor for interdecadal variations in the global monsoon (Wang et al. 2013; Deng et al. 2018). By influencing the interhemispheric thermal contrast and meridional energy transport, the decadal changes in North Atlantic SST anomalies can modulate the migration of the intertropical convergence zone (ITCZ) and Hadley circulation and further affect monsoon precipitation (T. Schneider et al. 2014).

Previous studies on global monsoon changes have mainly focused on the second half of the twentieth century. It has been found that the global land monsoon precipitation had an increasing trend during the period of 1901–55, which is opposite to that of the following half-century (Zhang and Zhou 2011). A statistical analysis between the precipitation and SST during the twentieth century suggests that the NHSM land precipitation change is highly correlated with the North Atlantic–tropical Indian Ocean meridional SST gradient index and the tropical Pacific zonal SST gradient index (Wang et al. 2018), but the processes dominating the precipitation changes and the circulation changes during the early half of the twentieth century remain unknown. This study investigates the forcing mechanism of global monsoon changes in the twentieth century by addressing the following questions: 1) How did global land monsoon precipitation and monsoon circulation change during the twentieth century? 2) Can the interdecadal variations in global monsoon be attributed to the atmospheric response to sea surface temperature changes? 3) Which oceanic region is the key driver of the interdecadal variations in the global monsoon?

To answer the above questions, we use the outputs of the extended AMIP simulations of two AGCMs driven by historical observed SST. Despite some limitations, especially the lack of two-way ocean–atmosphere coupling, this methodology has been shown to be favorable for understanding global and regional precipitation variability (Zhou et al. 2008b; Schubert et al. 2016; Deng et al. 2018; among many others). In addition, the single basin SST forced AGCM experiments from the U.S. CLIVAR drought working group are also used (Schubert et al. 2009).

The remainder of this paper is organized as follows. The datasets and methods used are described in [section 2](#). In [section 3](#), we first evaluate the performance of AGCMs in simulating global monsoon and then compare the changes in global monsoon precipitation between AGCMs and observation. The roles of SST changes in monsoon circulation and precipitation are also explored in [section 3](#). A summary and discussion are presented in [section 4](#).

2. Data and methods

a. Observational data

Three sets of monthly gauge-based precipitation datasets with a spatial resolution of $0.5^\circ \times 0.5^\circ$ are used as observational global land precipitation for the period of 1901–2000: 1) the precipitation dataset constructed by the Climatic Research Unit (CRU V4.01; [Harris and Jones 2017](#)), 2) the Full Data Reanalysis Product provided by the Global Precipitation Climatology Center (GPCC V7; [U. Schneider et al. 2014](#)), and 3) the precipitation dataset compiled by the University of Delaware (U. of Delaware V4.01; [Willmott and Matsuura 2000](#)). The discrepancies between different precipitation datasets in the estimation of historical precipitation changes over global land are pointed out by previous studies, which become apparent after 1991 and are mainly caused by the differences in data sources and interpolation algorithms ([Sheffield et al. 2012](#); [Trenberth et al. 2014](#); [Sun et al. 2018](#)). The principal sources for CRU are the World Meteorological Organization (WMO) and the National Climate Data Center (NCDC), while the U. of Delaware dataset is mainly based on the Global Historical Climatology Network-Monthly (GHCN-M) and Global Summary of the Day (GSOD) datasets ([Harris et al. 2014](#)). The CRU and GHCN together with other data sources are all integrated into the GPCC, making it own the largest basis of monthly precipitation data worldwide ([U. Schneider et al. 2014](#)). The GPCC's Full Data product includes monthly precipitation totals of more than 85 000 stations, while the CRU and U. of Delaware datasets have poor gauge coverage with more homogeneous and long time records ([U. Schneider et al. 2014](#); [Trenberth et al. 2014](#)). As both coverage and temporal continuity can affect the quality of precipitation datasets, we adopt the ensemble mean of these three datasets to reduce the uncertainties.

In addition, the ensemble mean of two sets of global monthly precipitation datasets for the period of 1979–2010 are used to define global monsoon domain: 1) the precipitation dataset provided by the Global Precipitation

Climatology Project (GPCP v2.2; [Adler et al. 2003](#)) and 2) the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP v1201; [Xie and Arkin 1997](#)). Both datasets are based on rain gauge data and satellite observations with a spatial resolution of $2.5^\circ \times 2.5^\circ$. The spatial patterns represented by these two products are significantly similar especially over land, while there is an artificial decreasing trend in oceanic precipitation for CMAP due to atoll sampling error ([Yin et al. 2004](#); [Zhou et al. 2008b](#)). The main difference for these two datasets over land is the use of the Television Infrared Observation Satellite (TIROS) Operational Vertical Sounder (TOVS) after July 1987 for GPCP. As we focus on global monsoon domain over land, both CMAP and GPCP are used to avoid the uncertainties due to different data sources and merging methods ([Yin et al. 2004](#); [Hao et al. 2016](#)).

In addition to precipitation datasets, the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST v1.1; [Rayner et al. 2003](#)), as a combination of the monthly global SST and sea ice concentration with a spatial resolution of $1^\circ \times 1^\circ$ from 1871 to the present, is used. The horizontal winds at 850 hPa derived from the Twentieth Century Reanalysis (20CR V2c; [Compo et al. 2011](#)) with a spatial resolution of $2^\circ \times 2^\circ$ are used to quantify whether AGCMs could reproduce climatological monsoon circulation. All the datasets are interpolated onto the same spatial resolution of $2.5^\circ \times 2.5^\circ$ by the bilinear interpolation method.

b. Model data

Two sets of AGCM simulations forced by historical observed SST are used:

- 1) Four realizations of the Met Office Hadley Centre model HadAM3 simulation driven by the HadISST and climate forcings over the period of 1869–2002. This model employs a horizontal resolution of 3.75° (longitude) $\times$ 2.5° (latitude). Data outputs at the surface, 850, 500, and 200 hPa are provided ([Pope et al. 2000](#)). The data are provided by the Climate Variability and Predictability (CLIVAR) C20C project ([Folland et al. 2002](#); [Scaife et al. 2009](#); [Zhou et al. 2009a](#)).
- 2) Six realizations of the National Aeronautics and Space Administration/Goddard Institute of Space Studies (NASA/GISS) model GISS-E2-R simulation driven by observed SST and sea ice over the period of 1900–2005. The data are downloaded from the CMIP5 archive ([Schmidt et al. 2014](#)). The model employs a horizontal resolution of 2.5° (longitude) $\times$ 2.0° (latitude) with 17 vertical levels.

For both AGCM simulations, ensemble methods are employed in which the integrations begin with different

reasonable atmospheric initial conditions but are forced by the same observed historical SST. The outputs are interpolated onto the same spatial resolution of $2.5^\circ \times 2.5^\circ$ by the bilinear interpolation method.

To isolate the role of specific basin's SST anomalies, the outputs of four sets of AGCM simulations forced by idealized SST anomaly patterns provided by the U.S. CLIVAR drought working group are also employed in this study. The associated forcing SST patterns are the three leading patterns of annual mean SST from 1901 to 2004 extracted by rotated empirical orthogonal function (REOF) based on HadISST, including a global warming pattern, a PDO-like pattern, and an AMO-like pattern (see Fig. 1 of Schubert et al. 2009). Scaled versions of the REOFs were then added to monthly varying SST climatology during the period of 1901–2004 to gain the full forcing patterns. The scaling factors are plus and minus two standard deviations of the associated PCs for the PDO-like and the AMO-like patterns, while they are plus and minus one standard deviation for the global warming pattern (Schubert et al. 2009).

In this study, we focus on seven baseline experiments (namely, PnAn, PcAn, PwAn, PnAc, PnAw, PcAw, and PwAc), which are the combinations of the PDO-like pattern and the AMO-like pattern with cold, neutral, or warm SST anomalies as shown in Table 1. The four AGCMs include the National Center for Atmospheric Research (NCAR) Community Climate Model version 3.0 (CCM3; Kiehl et al. 1998), the GFDL Atmosphere Model version 2.1 (GFDL AM2.1; Delworth et al. 2006), the NCAR Global Forecast System (GFS; Campana and Caplan 2005) and the NASA Seasonal-to-Interannual Prediction Project (NSIPP1; Bacmeister et al. 2000). All the simulations were integrated for 50 years and forced by the above SST forcing patterns except GFS (35 years). Atmospheric responses to different SST forcing patterns are obtained through the differences between the forced runs and the control run (i.e., PnAn). The climatology and the ensemble mean of the four AGCMs' outputs are used in this analysis. Six different forcing patterns (all except PnAn) are shown in Fig. S1 in the online supplemental material.

c. Analysis methods

The global monsoon domain is defined by the annual range of precipitation exceeding 2.0 mm day^{-1} and the local summer precipitation exceeding 55% of the annual precipitation based on the climatological mean of merged GPCP–CMAP data for the period of 1979–2010 (Wang et al. 2013). Annual range is defined as local summer precipitation minus local winter precipitation (Wang and Ding 2006). Local summer is defined from

TABLE 1. Seven baseline experiments used in this study provided by the U.S. CLIVAR Drought Working Group. In the experiment names, P represents a PDO-like pattern, A represents an AMO-like pattern, w (c) refers to the warm (cold) phase of the pattern scaled by plus two (minus two) standard deviations, and n refers to climatology (neutral).

	Cold Atlantic	Neutral Atlantic	Warm Atlantic
Cold Pacific	—	PcAn	PcAw
Neutral Pacific	PnAc	PnAn	PnAw
Warm Pacific	PwAc	PwAn	—

May to September (from November to March) for the Northern (Southern) Hemisphere, and local winter is defined from November to March (from May to September) for the Northern (Southern) Hemisphere (Wang et al. 2013).

To measure the strength of monsoon precipitation, the Northern Hemisphere Monsoon Index (NHMI), defined as summer precipitation averaged over the Northern Hemisphere monsoon domain, the Southern Hemisphere Monsoon Index (SHMI), defined as summer precipitation averaged over the Southern Hemisphere monsoon domain, and the global monsoon index (GMI), defined as the mean of NHMI and SHMI, are used (Wang and Ding 2006). In addition, the NHSM circulation intensity index and Walker circulation intensity index are used to measure the variability of monsoon circulation. The NHSM circulation intensity index is defined by the vertical shear of zonal winds between 850 and 200 hPa averaged in a strip from Mexico to the Philippines (0° – 20°N , 120°W – 120°E), and the Walker circulation intensity index is defined by the zonal winds at 850 hPa averaged over the equatorial Pacific (10°S – 10°N , 140°E – 120°W) (Wang et al. 2013).

A moisture budget analysis is used to obtain insights into the mechanisms responsible for the changes in precipitation (Seager et al. 2010; Chou and Lan 2012), which is written as

$$P = -\partial_t \langle q \rangle - \langle \nabla_h \cdot \mathbf{V}_h q \rangle - \langle \partial_p \omega q \rangle + E + \delta, \quad (1)$$

where P , E and q are precipitation, evaporation, and specific humidity, respectively. Also, $\mathbf{V}_h$ and ω denote horizontal wind and vertical velocity, respectively; δ is the residual term, which is attributed to subseasonal transient eddies, and angle brackets denote the mass integral from the surface to tropopause. As $-\partial_t \langle q \rangle$ is smaller compared with other terms on a monthly scale and $\omega \approx 0$ at the surface and tropopause, the moist budget equation can be simplified as

$$P = -\langle \mathbf{V}_h \nabla_h q \rangle - \langle \omega \partial_p q \rangle + E + \delta \quad (2)$$

and the change of precipitation can be decomposed as

$$P' = -\langle \mathbf{V}_h \nabla_h q \rangle' - \langle \omega \partial_p q \rangle' + E' + \delta', \quad (3)$$

where the prime denotes the monthly anomalies after removal of the climatological mean. The vertical moisture advection term $-\langle \omega \partial_p q \rangle'$ can be further divided as

$$-\langle \omega \partial_p q \rangle' = -\langle \bar{\omega} \partial_p q' \rangle - \langle \omega' \partial_p \bar{q} \rangle - \langle \omega' \partial_p q' \rangle, \quad (4)$$

and the first (second) term on the right-hand side of the equation is defined as the thermodynamic (dynamic) component of the vertical moisture advection term. The thermodynamic term reflects the contribution of specific humidity changes on precipitation, while the dynamic term reflects the contribution of atmospheric circulation changes on precipitation.

The empirical orthogonal function (EOF) method and the multivariate EOF (MV-EOF) method are used in our analysis. A 4-yr running average is used to remove interannual signals as in previous studies, and the average represents the value for the second year (van Oldenborgh et al. 2012; Wang et al. 2018). Sensitivity tests have been made with 5-, 9-, and 10-yr running averages; the results are essentially the same as those for the 4-yr running average, indicating that the decadal trends are not sensitive to the smoothing time windows (Fig. S2).

In our analysis, the linear trend is calculated by the least squares method with statistical significance tested by a 1000-ensemble Monte Carlo test (hereinafter MC test; Wilks 1995) and nonparametric Mann–Kendall test (hereinafter MK test; Kendall 1955). The statistical significance of correlation coefficients and regression coefficients is tested by the MC test. Because the moving average filter will lead to an artificial increase of autocorrelation, the time series are prewhitened before the MK test, and each sample for the MC test adopts the 4-yr running average.

3. Results

a. Evaluation of the simulation of the mean state of the global monsoon

To evaluate the performance of the two AGCMs in global monsoon (GM) simulations, the spatial patterns of climatological means for the annual range of global precipitation and 850-hPa winds during 1901–2000 derived from observations, 20CR, and simulations are shown in Fig. 1. In the observation, the areas where annual range exceeds 2 mm day^{-1} exist in the land monsoon domains over North America, South America, northern Africa, southern Africa, South Asia, East Asia,

and Australia (Fig. 1a, red lines). The seasonal reversal of 850-hPa winds features a strong cross-equatorial flow from East Africa and the western equatorial Indian Ocean to South Asia and from the Maritime Continent to the eastern Pacific. A strong equatorial westerly over the Atlantic is seen (Fig. 1a).

Both AGCMs reasonably reproduce the climatology for the annual range of global land precipitation and low-level winds (Figs. 1b,c). The monsoon rainbands and major rainfall centers are well simulated. The pattern correlation coefficient (PCC) between the simulation and the observation for the annual range of global land monsoon is 0.894 for HadAM3 and 0.773 for GISS-E2-R, both of which are statistically significant at the 5% level by *t* test. We note that the annual range for GISS-E2-R is weaker compared with the observations and HadAM3, especially over South Asia (Fig. 1b), primarily due to the dry biases during summer (figure not shown). Summer precipitation averaged over the land monsoon domain reaches 6.08 mm day^{-1} for observation, while it is 6.47 and 5.61 mm day^{-1} for HadAM3 and GISS-E2-R, respectively. The observed cross-equatorial flows over the Indian and Pacific Oceans and equatorial westerly over the Atlantic are evident in both AGCM simulations. The reasonable performance of the two AGCMs in reproducing the climatology of monsoon rainfall and monsoon circulation adds confidence to our following analysis of decadal changes in GM.

b. Interdecadal variability of GM precipitation over land

The time series of the GMI, NHMI, and SHMI and the linear trends of these three indices during 1901–55 and 1956–90 are shown in Fig. 2. As in Zhang and Zhou (2011), the GMI and NHMI in observation show significant increasing trends (0.051 and $0.053 \text{ mm day}^{-1} \text{ decade}^{-1}$, respectively) during 1901–55 and significant decreasing trends (-0.056 and $-0.121 \text{ mm day}^{-1} \text{ decade}^{-1}$, respectively) during 1956–90. The SHMI shows a weaker increasing trend ($0.049 \text{ mm day}^{-1} \text{ decade}^{-1}$) from 1901 to 1955, but that is not followed by a decreasing trend. The interdecadal variations in GMI are mainly contributed by the NHMI as previously reported (Zhang and Zhou 2011), which may be due to the larger fraction of land in the Northern Hemisphere, and the thermal contrast in the two hemispheres is favorable for the intensification of the NHSM (Wang et al. 2012). The linear trends in the two periods are robust, as evidenced by the running linear trends (Fig. 2e).

The significant increasing trends of the three monsoon indices during 1901–55 can be reasonably reproduced by HadAM3 (Fig. 2d, Table 2). The simulated increasing trends of GMI ($0.014 \text{ mm day}^{-1} \text{ decade}^{-1}$) and NHMI

FIG. 1. Climatological mean for the annual range of global precipitation (mm day^{-1} ; shading) and 850-hPa winds (m s^{-1} ; vectors). The annual range is defined by the local summer mean minus the local winter mean. The climatology is derived from the period of 1901–2000. The wind vectors shown exceed 2 m s^{-1} . The red lines indicate the boundaries of the global land monsoon domain based on merged GPCP-CMAP data. (a) Observation and 20CR, (b) GISS-E2-R, and (c) HadAM3.

($0.035 \text{ mm day}^{-1} \text{ decade}^{-1}$) for GISS-E2-R are weaker than observation and HadAM3, and an insignificant decreasing trend in SHMI is even seen. For the period of 1956–90, both AGCMs reasonably reproduce the significant decreasing trends in NHMI (-0.113 and $-0.072 \text{ mm day}^{-1} \text{ decade}^{-1}$ for GISS-E2-R and HadAM3, respectively). The decreasing trend of GMI ($-0.053 \text{ mm day}^{-1} \text{ decade}^{-1}$) is weakly reproduced by GISS-E2-R and is insignificant.

The above results indicate that changes in the global land monsoon precipitation throughout the twentieth century are mainly contributed by the Northern Hemisphere. Therefore, we focus on the changes in NHSM precipitation over land in subsequent analysis. To highlight the decadal variability of NHSM precipitation, we further perform EOF analysis. The spatial patterns

for the first leading EOF mode of NHSM precipitation derived from the observation and the AGCM simulations are shown in Fig. 3. The variance explained by the first leading EOF mode is 12.85% in observation. The majority of the Northern Hemisphere land monsoon region shows coherent positive precipitation anomalies with the maximum in northern Africa (Fig. 3a). We note the patchy signals over the Asian monsoon domain, which are also evident in the spatial patterns of the trends of the NHSM precipitation during the two periods (Fig. S3). The simulated patterns are similar to that for observation, but with some biases mainly located in the Asian monsoon domain. For example, evident biases are seen over south India and the Indo-China Peninsula in the results of HadAM3 (Fig. 3c). The explained

FIG. 2. The time series (mm day^{-1}) of (a) the global monsoon index, (b) the Northern Hemisphere monsoon index, and (c) the Southern Hemisphere monsoon index derived from observation (black), GISS-E2-R (blue), and HadAM3 (red). (d) The linear trends ($\text{mm day}^{-1} \text{ decade}^{-1}$) of the three indices for the periods of 1901–55 and 1956–90. The thick lines in (a)–(c) and circles in (d) are the ensemble-mean results. The light shades in (a)–(c) and error bars in (d) are for plus and minus one standard deviation about the ensemble mean, which are calculated from six and four members of GISS-E2-R and HadAM3, respectively. The solid circles in (d) indicate the linear trends are significant at the 5% level both by MC and MK tests. (e) The running linear trends for NHMI derived from observation. Each column represents the start year, and each row represents the end year used for calculating the trend. The stippling indicates that the trends are significant at the 5% level using the MK test. The blue dots represent the two periods (1901–55 and 1956–90).

variances of the first leading EOF mode for simulations are higher than that for observation, 35.73% and 46.48% for GISS-E2-R and HadAM3, respectively.

The corresponding time series of the first leading EOF mode of NHSM land precipitation (PC1) and the trends of PC1 during the two periods are shown in Fig. 4. The simulations show coherent variations with the observations. All the time series show increasing trends during 1901–55 and decreasing trends during 1956–90, such as the monsoon indices, with the linear trends being statistically significant at the 5% level by the MK and MC tests.

In summary, the global land monsoon precipitation strengthens during 1901–55 and then weakens, which is mainly contributed by the NHSM land precipitation coherent changes. Both AGCM simulations well reproduce the interdecadal variability in global land monsoon precipitation and NHSM land precipitation. We investigate the main circulation characteristics related

to NHSM land precipitation changes in the subsequent section.

c. Circulation characteristics related to GM precipitation changes

The global monsoon system is a global-scale overturning of atmosphere with the three-dimensional spatial structure varying with seasonal changes. The complete structure of the global monsoon includes divergence in the upper troposphere with a maximum at 150 hPa, convergence in the lower troposphere, which reaches a maximum at 925 or 850 hPa based on different datasets and accompanying vertical motion (Trenberth et al. 2000). To reveal the main circulation characteristics related to GM precipitation changes, the time series of the NHSM circulation intensity indices and the Walker circulation intensity indices derived from the two AGCMs are shown in Fig. 5. The NHSM circulation intensity index reflects the changes in both low-level and

TABLE 2. The trends ($\text{mm day}^{-1} \text{ decade}^{-1}$) calculated based on the least squares method and correlated p value (%) determined using MK and MC tests for GMI, NHMI, and SHMI during two periods.

	Trend		MK		MC	
	1901–55	1956–90	1901–55	1956–90	1901–55	1956–90
GMI						
OBS-ens	0.051	−0.056	1	5	1	—
GISS-E2-R	0.014	−0.053	—	1	—	—
HadAM3	0.069	−0.064	1	1	1	—
NHMI						
OBS-ens	0.053	−0.121	1	1	1	1
GISS-E2-R	0.035	−0.113	1	1	1	1
HadAM3	0.073	−0.072	1	1	1	5
SHMI						
OBS-ens	0.049	0.009	1	—	1	—
GISS-E2-R	−0.006	0.006	—	—	—	—
HadAM3	0.064	−0.056	1	5	1	—

upper-level zonal winds over the Northern Hemisphere monsoon domain (Wang et al. 2013). The Walker circulation intensity index reflects the changes in equatorial easterlies over the equatorial Pacific and are tightly related to the NHSM system (Trenberth et al. 2000). The NHSM circulation intensity indices are highly

correlated with PC1 ($r = 0.924$ and 0.808 for GISS-E2-R and HadAM3, respectively; $p < 0.05$). The Walker circulation intensity indices are highly negatively correlated with PC1 ($r = -0.332$ and -0.461 for GISS-E2-R and HadAM3, respectively; $p < 0.05$). Accompanied by the increase of NHSM land precipitation as represented

FIG. 3. The spatial pattern of the first leading EOF mode of summer precipitation over the Northern Hemisphere land monsoon domain. The red lines represent the boundaries of the Northern Hemisphere land monsoon domain. (a) Observation, (b) GISS-E2-R, and (c) HadAM3.

FIG. 4. (a) The corresponding time series of the first leading EOF mode (PC1) of summer precipitation over the Northern Hemisphere land monsoon domain. (b) The linear trends of PC1 for the periods of 1901–55 and 1956–90. The linear trends of PC1 derived from different datasets during the two periods are all significant at the 5% level using the MK and MC tests.

by PC1, there is a strengthening of equatorial easterlies at the lower troposphere located in the equatorial western Pacific (Fig. 5b), which can strengthen the East Asian and Australian summer monsoon. The associated strengthened Walker circulation increases the moisture convergence over the Maritime Continent. The westerly anomalies at the lower troposphere in the equatorial eastern Pacific, equatorial Atlantic, and equatorial Indian Ocean (Fig. 5a) strengthen the North American, North African and Asian summer monsoon circulation and lead to the increase in monsoon precipitation in these monsoon regions.

Can the changes in circulation explain the interdecadal variability of NHSM land precipitation? We further perform a moist budget analysis to reveal the physics behind the wetting and drying trends of NHSM land precipitation during the two different time periods. The linear trends of area-averaged moisture budget components of NHSM land precipitation derived from GISS-E2-R are shown in Fig. 6. During 1901–55, the vertical moisture advection term shows a significant upward trend along with coherent variations in evaporation (Fig. 6a). The increasing trend in precipitation is dominated by the vertical moisture advection term. The

vertical moisture advection term could be further separated into a thermodynamic component due to the changes in specific humidity, a dynamic component due to the changes in atmospheric circulation, and a non-linear component. Both the thermodynamic component and the dynamic component contribute to the wetting NHSM, and the value is larger for the thermodynamic component, while the increasing trend is insignificant for the area-averaged dynamic component.

Focusing on the spatial patterns, a close resemblance is seen between the precipitation, the vertical moisture advection term, and the dynamic component, while the thermodynamic component contributes to a coincident wetting trend over the NHSM domain (Figs. 7 and 8; see also Fig. S4, left column). Although the monsoon precipitation shows wetting trends over most of the monsoon domains, there are still some areas with opposite trends, such as southwest China (Fig. 7a) and the southern part of the northern African (NAF) monsoon domain (Fig. 8a). This phenomenon can be explained by the dynamic component, but the areas with decreasing trends are larger for the dynamic component, especially for the NAF monsoon domain (Fig. 8g), leading to an insignificant decreasing trend ($-0.004 \text{ mm day}^{-1} \text{ decade}^{-1}$; Table 3) for the area-averaged

FIG. 5. The corresponding time series of the first leading EOF mode (PC1) of summer precipitation over the Northern Hemisphere land monsoon domain derived from the ensemble mean of the two models (black) and (a) the Northern Hemisphere monsoon circulation intensity indices (m s^{-1}) and (b) the Walker circulation intensity indices (m s^{-1}) derived from GISS-E2-R (blue) and HadAM3 (red). The thick lines are the ensemble-mean results. The light shades are for plus and minus one standard deviation about the ensemble means, which are calculated from six and four members of GISS-E2-R and HadAM3, respectively. The numbers in the top-right corners are the correlation coefficients between the circulation intensity indices and PC1.

dynamic component over this monsoon domain. For the Asian monsoon domain, the dynamic ($0.029 \text{ mm day}^{-1} \text{ decade}^{-1}$) and thermodynamic components ($0.033 \text{ mm day}^{-1} \text{ decade}^{-1}$) are equally important, while the dynamic component ($0.144 \text{ mm day}^{-1} \text{ decade}^{-1}$) is the major contributor of the wetting trend of monsoon precipitation ($0.217 \text{ mm day}^{-1} \text{ decade}^{-1}$) over the North American (NAM) monsoon domain.

For the period of 1956–90, the vertical moisture advection term shows a significant downward trend and contributes to the drying trend of monsoon precipitation (Fig. 6b). The change in the vertical moisture advection term is mainly attributed to the decreasing dynamic component as the thermodynamic component has no significant trend. The spatial patterns of corresponding moisture budget terms also reveal the prominent role of the dynamic term related to atmospheric circulation changes (Figs. 7 and 8; see also Fig. S4, right column).

The results of three individual monsoon domains are similar to that for the Northern Hemisphere land monsoon domain. The significant downward trends of vertical moisture advection term (-0.332 , -0.090 , and $-0.177 \text{ mm day}^{-1} \text{ decade}^{-1}$ for NAM, NAF, and Asia, respectively; $p < 0.05$), which are dominated by the changes in the dynamic components (-0.276 , -0.088 , and $-0.063 \text{ mm day}^{-1} \text{ decade}^{-1}$ for NAM, NAF, and Asia, respectively; $p < 0.05$), have led to the drying trends of monsoon precipitation in three individual monsoon domains (-0.270 , -0.174 , and $-0.174 \text{ mm day}^{-1} \text{ decade}^{-1}$ for NAM, NAF, and Asia, respectively; $p < 0.05$). The contribution of thermodynamic components for three individual monsoon domains can be neglected compared with the dynamic components during 1956–90. Thus, the drying trend in the recent decades differs with the wetting trend during 1901–55 in the contribution of thermodynamic

FIG. 6. The linear trends ($\text{mm day}^{-1} \text{ decade}^{-1}$) of moisture budget terms averaged over the Northern Hemisphere land monsoon domain for the periods of (a) 1901–55 and (b) 1956–90. The stars indicate the linear trends are significant at the 5% level both using the MC and MK tests. The data used are derived from GISS-E2-R.

components. Opposite trends are seen over the Indo-china Peninsula (Figs. 7b,d,h).

The above results indicate that the interdecadal variability in NHSM land precipitation is mainly contributed by the dynamic component of the vertical moisture advection term, while the thermodynamic component leads to a coincident wetting trend over the NHSM domain during both periods. In other words, the changes in atmosphere circulation and accompanying vertical velocity can explain the precipitation variability in the twentieth century. We further discuss the forcing mechanisms of NHSM circulation changes in the subsequent section.

d. Forcing mechanisms of monsoon circulation changes

Previous studies have highlighted the role of tropical Pacific and North Atlantic SST in driving the NHSM system since the late 1970s (Wang et al. 2013). A later extended analysis emphasized the meridional thermal contrast between the North Atlantic and South Indian Oceans and the zonal SST gradient across the tropical Pacific (Wang et al. 2018). To reveal which parts of anomalous global SST dominate the changes in the NHSM system during the twentieth century, MV-EOF analysis is applied to decompose the covariability between NHSM precipitation and global SST in summer, and the three leading MV-EOF modes are shown in Fig. 9. The first MV-EOF mode between NHSM precipitation and SST accounts for 36.41% of the total covariance. Associated with the first mode, significantly

warming SST anomalies in the global scope and a corresponding drying trend over the Northern Hemisphere monsoon domain can be seen, reflecting the influence of global warming on monsoon precipitation. The second and third MV-EOFs explain, respectively, 12.48% and 7.91% of the total covariance. The associated SST anomalies show a negative PDO-like pattern (Fig. 9h) and a positive AMO-like pattern (Fig. 9i). Both SST patterns are accompanied by wetting NHSM precipitation, and the corresponding time series show interdecadal variations with turning points between 1940 and 1960.

To examine whether the combination of anomalies associated with the negative PDO and the positive AMO lead to the interdecadal variations in NHSM precipitation, a combined AMO–PDO index is defined as

$$\text{AMO–PDO index} = 0.353\text{PC2} + 0.281\text{PC3},$$

and the correlation coefficient between the combined AMO–PDO index and the corresponding time series for the first EOF mode of NHSM precipitation reaches 0.710 ($p < 0.05$), while it is 0.449 and 0.577 for PC2 and PC3, respectively.

The associated SST patterns for the three leading MV-EOF modes resemble the three leading patterns of annual mean SST from 1901 to 2004 extracted by REOF (see Fig. 1 of Schubert et al. 2009). We further use the outputs of the AGCM simulations forced by idealized SST anomaly patterns provided by the U.S. CLIVAR drought working group to further investigate the

FIG. 7. The spatial patterns of the trends ($\text{mm day}^{-1} \text{ decade}^{-1}$) of (a),(b) precipitation, (c),(d) the vertical moisture advection term, and (e),(f) the thermodynamic and (g),(h) the dynamic terms of the vertical moisture advection term during (left) 1901–55 and (right) 1956–90 over the Asian monsoon domain. The blue diagonals indicate that the linear trends are significant at the 5% level using the MK test. The purple lines represent the boundaries of the land monsoon domain. The data used are derived from GISS-E2-R.

contribution of the PDO and AMO on NHSM precipitation. Responses of area-averaged summer precipitation over the Northern Hemisphere monsoon domain to six different forcing patterns are shown in Fig. 10. A high consistency is seen in the precipitation

responses for the four AGCMs, although GFS (GFDL) shows a slightly different signal in the PnAw (PwAn) experiment. Compared with PnAn, the NHSM precipitation is significantly increased for PnAw ($0.108 \text{ mm day}^{-1}$ for the ensemble mean), while it is significantly less for PnAc

FIG. 8. As in Fig. 7, but for the northern African land monsoon domain.

($-0.120 \text{ mm day}^{-1}$ for the ensemble mean), indicating that the warm phase of the North Atlantic accompanied by a neutral condition in the Pacific is favorable for a wetting NHSM.

In addition, the cold phase of the tropical Pacific is also favorable for a wetting NHSM, as evidenced by more precipitation in the PcAn experiment ($0.114 \text{ mm day}^{-1}$ for the ensemble mean) and less precipitation ($-0.167 \text{ mm day}^{-1}$

for the ensemble mean) in the PwAn experiment compared with PnAn. The contributions of positive AMO and negative PDO are comparable.

We also compared the results of PcAw ($0.184 \text{ mm day}^{-1}$ for the ensemble mean) and PwAc ($-0.257 \text{ mm day}^{-1}$ for the ensemble mean) with PnAn and find that following a warmer North Atlantic and colder tropical Pacific, the NHSM precipitation would enhance (and vice versa);

TABLE 3. The linear trends ($\text{mm day}^{-1} \text{decade}^{-1}$) of moisture budget (MB) terms averaged over the North American (NAM), northern African (NAF), and Asian land monsoon domains for the periods of 1901–55 and 1956–90. One and two asterisks indicate that the linear trends are significant at the 1% and 5% level, respectively, using the MC test. The data used are derived from GISS-E2-R.

MB term	1901–55			1956–90		
	NAM	NAF	Asia	NAM	NAF	Asia
P	0.217*	0.061*	0.098*	−0.270*	−0.174*	−0.174**
E	0.048*	0.020*	0.024*	−0.103*	−0.087*	−0.049*
$\langle udq \rangle$	0.047*	0.009*	0.012*	−0.044*	0.011	−0.005
$\langle vdq \rangle$	−0.026*	0.002	0.006	0.020	−0.006	−0.027
$\langle wdq \rangle$	0.185*	0.007	0.117*	−0.332*	−0.090*	−0.177**
$\langle wdq' \rangle$	0.014*	0.015*	0.033*	−0.006	0.006	−0.023
$\langle w'dq \rangle$	0.144*	−0.004	0.029*	−0.276*	−0.088*	−0.063**

this kind of response is also evident in the spatial patterns of the ensemble mean (Fig. S5). Therefore, a combination of AMO in positive phase and PDO in negative phase is favorable for enhanced NHSM precipitation and vice versa. The model responses also aid in understanding why the combined AMO–PDO index shown above explains more variance than either the AMO or PDO index alone in observation.

To understand how SST anomalies drive the NHSM, the tropospheric mean temperature indicated by the boreal summer mid-to-upper tropospheric thickness, which is defined as the difference of geopotential height between 200 and 500 hPa, and 200-hPa wind anomalies are regressed onto the corresponding time series of the

second and third leading MV-EOF modes, respectively (Fig. 11). Based on the thermal wind balance theory, vertical wind shear is balanced with the horizontal gradient of tropospheric atmosphere thickness, which is determined by tropospheric temperature gradient. Accompanied by the intensification of the PC2, the mid-to-upper-tropospheric thickness features a significant decrease over the tropical central to eastern Pacific and increases over the subtropical Pacific, Atlantic, Africa, and Eurasia (Figs. 11e,f). The pattern strongly resembles the Gill–Matsuno response to anomalous tropical Pacific SST forcing with two strong off-equatorial Rossby wave maxima and a strong equatorial Kelvin wave maximum (Matsuno 1966; Gill 1980). The negative

FIG. 9. MV-EOF analysis between NHSM precipitation and SST during summer for the period of 1901–2000. (a)–(c) The corresponding time series for three leading modes (black lines), (d)–(f) the three leading patterns for precipitation, and (g)–(i) the three leading patterns for SST. The red lines in (b) and (c) represent the corresponding time series for the first EOF mode of NHSM precipitation, and the blue lines in (b) and (c) represent the AMO–PDO index.

FIG. 10. Difference in the area-averaged summer precipitation (mm day^{-1}) over the global monsoon domain in six experiments from the control run (PnAn) for four AGCMs and the ensemble mean.

phase of PDO drives an increased meridional tropospheric temperature gradient; this further leads to anomalous equatorial east winds and a strengthened tropical easterly jet, which can influence the intensity of monsoon precipitation over the Afro-Asian domain (Feudale and Kucharski 2013). Meanwhile, low sea level pressure is seen over North America, the North Atlantic, the Saharan region, and central Asia in both AGCMs (Figs. 12e,f). The low pressure over these regions is favorable for the convergence at low level and the strengthening of the westerlies around the south edge, which transport more water vapor into the NAM, NAF, and SAS monsoon domains and lead to the increase of monsoon precipitation.

Turning to the PC3, forced by the positive phase of AMO, the warming North Atlantic drives a warming troposphere as indicated by the positive geopotential thickness at extratropical regions, leading to a meridional tropospheric temperature gradient, which is favorable for the strengthening of the tropical easterly at upper level (Figs. 11c,d). Low sea level pressure over North America, the North Atlantic, the Saharan region, and central Asia and associated westerlies are also evident following the warming North Atlantic. A similar analysis is applied to the NHSM circulation intensity index (Figs. 11a,b and 12a,b). The spatial pattern is close to the combination of anomalies associated with positive AMO and negative PDO, demonstrating the effective driving of the interdecadal Pacific and North Atlantic SST changes to the NHSM circulation.

The two AGCMs show divergent responses in sea level pressure in the East Asian (EAS) monsoon domain and the northwestern Pacific (Fig. 12). Compared with other regional monsoons, the skill of monsoon precipitation is lower in the Asian–Australian monsoon

domain (Fig. 3), which is partly due to the neglect of air–sea interaction (Krishnamurthy and Kirtman 2003; Wang et al. 2005; Zhou et al. 2009b; Deng et al. 2018). Through the inclusion of air–sea interaction (Zou and Zhou 2013, 2016; Zou et al. 2016; Zou et al. 2018; Jiang et al. 2019), better reproduction of interdecadal and long-term changes in precipitation over the Asian–Australian monsoon domain is hoped for, which is a target of the CMIP6 Global Monsoon Modeling Intercomparison Project (Zhou et al. 2016).

In summary, a combination of AMO in positive phase and PDO in negative phase is favorable for the increase in NHSM precipitation. Both SST anomaly patterns can lead to a meridional gradient in mid-to-upper-tropospheric thickness between equatorial regions and extratropical regions and further strengthen the equatorial easterlies at upper levels. In addition, low sea level pressure is seen over North America, the North Atlantic, the Saharan region, and central Asia, leading to the strengthening of the westerlies and the increase in monsoon precipitation over NAF, NAM, and South Asia (SAS). How the North Atlantic and Pacific modulate the interdecadal variations in EAS monsoon precipitation remains inconclusive, since the two AGCMs show divergent responses and neither shows signals consistent with the observation.

4. Summary and concluding remarks

Global land monsoon precipitation underwent an increasing trend during the 1901–55 that was followed by a weakening trend up to 1990. Mechanisms behind the observed multidecadal variations in monsoon precipitation are addressed by analyzing the outputs of two AGCM ensemble simulations driven by the historical sea surface

FIG. 11. Summer geopotential thickness between 200 and 500 hPa and the 200-hPa wind anomalies regressed onto (a),(b) the Northern Hemisphere summer monsoon circulation index, and the corresponding time series of (c),(d) the third and (e),(f) the second leading MV-EOF modes. Plots are derived from (left) GISS-E2-R and (right) HadAM3. The stippling indicates the regression coefficients are significant at the 5% level using the MC test. The purple lines represent the boundaries of the Northern Hemisphere land monsoon domain.

temperature covering the twentieth century, supplemented with the single basin idealized SST forced AGCM experiments. The major results are summarized below.

The two AGCM simulations can well reproduce the climatological characteristics of global monsoon circulation and precipitation over global land monsoon domains. The observed monsoon rainbands and rainfall centers over North America, South America, northern Africa, southern Africa, South Asia, East Asia, and Australia can be reasonably simulated by the two AGCMs. The observed significant increasing trend in monsoon precipitation during 1901–55 and the decreasing trend during 1956–90 are mainly contributed by the NHSM land precipitation changes. The two AGCM simulations can reasonably reproduce the interdecadal variations in global land monsoon precipitation and NHSM land precipitation. Based on EOF analysis, the leading mode of NHSM land precipitation indicates coherent changes over the Northern Hemisphere land monsoon domain. The corresponding PCs show increasing

trends during 1901–55 and decreasing trends during 1956–90, such as the monsoon indices for both the observations and the two AGCMs.

Moisture budget analyses indicate that the increasing NHSM land precipitation during 1901–55 is contributed by the thermodynamic term associated with humidity changes and the dynamic component of vertical moisture advection term associated with the atmospheric circulation changes. The former leads to coincident wetting over the NHSM domain, while the latter determines the spatial patterns of the precipitation trends. However, the decreasing NHSM land precipitation during 1956–90 is mainly contributed by the dynamic component of the vertical moisture advection term, as the thermodynamic term has no significant trend for the same period. The increase (decrease) of NHSM land precipitation during the period of 1901–55 (1956–90) is associated with the strengthening (weakening) of the NHSM circulation, the Walker circulation, and accompanying vertical velocity.

FIG. 12. Summer sea level pressure and 850-hPa wind anomalies regressed onto (a),(b) the Northern Hemisphere summer monsoon circulation index, and the corresponding time series of (c),(d) the third and (e),(f) the second leading MV-EOF modes. Plots are derived from (left) GISS-E2-R and (right) HadAM3. The stippling indicates the regression coefficients are significant at the 5% level using the MC test. The purple lines represent the boundaries of the Northern Hemisphere land monsoon domain.

The interdecadal variations in global monsoon circulation can be modulated by the AMO and PDO. In particular, a combination of AMO in positive phase and PDO in negative phase is favorable for a stronger NHSM and vice versa. The warmer North Atlantic together with the colder eastern tropical Pacific and warmer western subtropical Pacific lead to a meridional gradient in geopotential thickness between equatorial regions and monsoon domains at upper-level and low sea level pressure over North America, the North Atlantic, the Saharan region, and central Asia. The pressure gradient could further strengthen summer monsoons, which transport more water vapor into monsoon regions and thereby drive the increase in monsoon precipitation.

In this study, the Northern Hemisphere monsoon domain is regarded as an integrated system to understand the mechanism of decadal variability. While this strategy is helpful for revealing the gross features of all regional monsoons in the context of long-term variability, attention should also be paid to the specific regional characteristics of monsoon variability. For example, an east–west dipole

mode of South Asian monsoon precipitation has been identified between central India and Bangladesh/Myanmar, which is modulated by different phases of El Niño (Yang et al. 2015). At an interdecadal scale, the warm (cold) phase of the PDO can also lead to dryness (wetness) in India and wetness (dryness) in Bangladesh/Myanmar (see Figs. 4 and 5 of Krishnamurthy and Krishnamurthy 2014). In addition, there is a 40-yr period oscillation for the main monsoon precipitation belt in East Asia (Ding et al. 2018). A “southern drought and northern flood” rainfall pattern could be found over East Asia during the 1950s and 1970s, while this pattern has been reversed since the late 1970s, and the rainband moved north in the early 1990s (Yu et al. 2004; Yu and Zhou 2007; Zhou et al. 2009c; Zhu et al. 2011; Ding et al. 2018). The PDO is the prime driver for the interdecadal shifts of the East Asian monsoon (Li et al. 2010; Qian and Zhou 2014; Zhu et al. 2015). The regional difference in monsoon precipitation may result in patchy signals over the Asian monsoon domain for EOF analysis (Fig. 3) and linear trends (Fig. S3 and Fig. 7).

Finally, we acknowledge that this study focuses on multidecadal variability in NH summer monsoons. While the results demonstrate the essential roles of natural SST variability in the Pacific and North Atlantic in modulating the interdecadal monsoon variability, we should not preclude the impacts of external forcings. Numerical simulations have shown evidence that anthropogenic aerosols can weaken the thermal contrast between land and ocean and then weaken monsoon circulation through affecting the radiation budget and the properties of clouds (Ramanathan et al. 2001; Li et al. 2011). The drying trend of monsoon precipitation over the land monsoon regions of South Asia (Bollasina et al. 2011), East Asia (Song et al. 2014), Africa (Held et al. 2005), and the Northern Hemisphere after the 1950s (Polson et al. 2014) can be partly explained by anthropogenic aerosols. In the meantime, the increasing greenhouse gases are favorable for increasing monsoon precipitation (Kitoh et al. 2013; Song et al. 2014; Chen and Zhou 2015). Further efforts should be devoted to understanding and distinguishing the externally forced and internally driven changes in monsoon precipitation, which is also one of the tasks of the CMIP6 Global Monsoon Modeling Intercomparison Project (Zhou et al. 2016).

Acknowledgments. This work is jointly supported by Chinese Academy of Sciences (Grants XDA20060102 and 134111KYSB20160031), and the National Natural Science Foundation of China (Grants 41775091 and 41330423). We appreciate the model data availability from CMIP5 and the CLIVAR C20C project. We also acknowledge the support from Jiangsu Collaborative Innovation Center for Climate Change.

REFERENCES

- Adler, R. F., and Coauthors, 2003: The version-2 Global Precipitation Climatology Project (GPCP) monthly precipitation analysis (1979–present). *J. Hydrometeorol.*, **4**, 1147–1167, [https://doi.org/10.1175/1525-7541\(2003\)004<1147:TVGPCP>2.0.CO;2](https://doi.org/10.1175/1525-7541(2003)004<1147:TVGPCP>2.0.CO;2).
- Arias, P. A., R. Fu, and K. C. Mo, 2012: Decadal variation of rainfall seasonality in the North American monsoon region and its potential causes. *J. Climate*, **25**, 4258–4274, <https://doi.org/10.1175/JCLI-D-11-00140.1>.
- Bacmeister, J., P. J. Pegion, S. D. Schubert, and M. J. Suarez, 2000: An atlas of seasonal means simulated by the NSIPP 1 atmospheric GCM. NASA Tech. Memo. 104606, Vol. 17, Goddard Space Flight Center, 194 pp.
- Bollasina, M. A., Y. Ming, and V. Ramaswamy, 2011: Anthropogenic aerosols and the weakening of the South Asian summer monsoon. *Science*, **334**, 502–505, <https://doi.org/10.1126/science.1204994>.
- Campana, K., and P. Caplan, Eds., 2005: Technical procedure bulletin for the T382 Global Forecast System. NOAA/NCEP/EMC, http://www.emc.ncep.noaa.gov/gc_wmb/Documentation/TPBoct05/T382.TPB.FINAL.htm.
- Chen, X., and T. Zhou, 2015: Distinct effects of global mean warming and regional sea surface warming pattern on projected uncertainty in the South Asian summer monsoon. *Geophys. Res. Lett.*, **42**, 9433–9439, <https://doi.org/10.1002/2015GL066384>.
- Chou, C., and C. W. Lan, 2012: Changes in the annual range of precipitation under global warming. *J. Climate*, **25**, 222–235, <https://doi.org/10.1175/JCLI-D-11-00097.1>.
- Compo, G. P., and Coauthors, 2011: The Twentieth Century Reanalysis Project. *Quart. J. Roy. Meteor. Soc.*, **137**, 1–28, <https://doi.org/10.1002/qj.776>.
- Delworth, T. L., and Coauthors, 2006: GFDL's CM2 global coupled climate models. Part I: Formulation and simulation characteristics. *J. Climate*, **19**, 643–674, <https://doi.org/10.1175/JCLI3629.1>.
- Deng, K., S. Yang, M. Ting, Y. Tan, and S. He, 2018: Global monsoon precipitation: Trends, leading modes, and associated drought and heat wave in the Northern Hemisphere. *J. Climate*, **31**, 6947–6966, <https://doi.org/10.1175/JCLI-D-17-0569.1>.
- Ding, Y., D. Si, Y. Liu, Z. Wang, Y. Li, L. Zhao, and Y. Song, 2018: On the characteristics, driving forces and inter-decadal variability of the East Asian summer monsoon (in Chinese). *Chin. J. Atmos. Sci.*, **42**, 533–558.
- Feudale, L., and F. Kucharski, 2013: A common mode of variability of African and Indian monsoon rainfall at decadal timescale. *Climate Dyn.*, **41**, 243–254, <https://doi.org/10.1007/s00382-013-1827-4>.
- Folland, C., J. Shukla, J. Kinter, and M. Rodwell, 2002: The climate of the twentieth century project. *CLIVAR Exchanges*, No. 7, International CLIVAR Project Office, Southampton, United Kingdom, 37–39.
- Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.*, **106**, 447–462, <https://doi.org/10.1002/qj.49710644905>.
- Goswami, B. N., M. S. Madhusoodanan, C. P. Neema, and D. Sengupta, 2006: A physical mechanism for North Atlantic SST influence on the Indian summer monsoon. *Geophys. Res. Lett.*, **33**, L02706, <https://doi.org/10.1029/2005GL024803>.
- Grimm, A. M., and J. P. J. Saboia, 2015: Interdecadal variability of the South American precipitation in the monsoon season. *J. Climate*, **28**, 755–775, <https://doi.org/10.1175/JCLI-D-14-00046.1>.
- Hao, Y., C. Zhu, and B. Liu, 2016: Discrepancies in boreal summer monsoon rainfall between GPCP and CMAP products during 1979–2014. *Atmos. Oceanic Sci. Lett.*, **9**, 226–233, <https://doi.org/10.1080/16742834.2016.1168603>.
- Harris, I., and P. D. Jones, 2017: CRU TS4.01: Climatic Research Unit (CRU) Time-Series (TS) version 4.01 of high-resolution gridded data of month-by-month variation in climate (Jan. 1901–Dec. 2016), accessed 28 November 2018, <https://doi.org/10.5285/58a8802721c94c66ae45c3baa4d814d0>.
- , —, T. J. Osborn, and D. H. Lister, 2014: Updated high-resolution grids of monthly climatic observations—The CRU TS3.10 dataset. *Int. J. Climatol.*, **34**, 623–642, <https://doi.org/10.1002/joc.3711>.
- Held, I. M., T. L. Delworth, J. Lu, K. L. Findell, and T. R. Knutson, 2005: Simulation of Sahel drought in the 20th and 21st centuries. *Proc. Natl. Acad. Sci. USA*, **102**, 17 891–17 896, <https://doi.org/10.1073/pnas.0509057102>.
- Hoerling, M., J. Hurrell, J. Eischeid, and A. Phillips, 2006: Detection and attribution of twentieth-century northern and southern African rainfall change. *Bull. Amer. Meteor. Soc.*, **19**, 3989–4008, <https://doi.org/10.1175/JCLI3842.1>.

- Hsu, P. C., T. Li, and B. Wang, 2011: Trends in global monsoon area and precipitation over the past 30 years. *Geophys. Res. Lett.*, **38**, L08701, <https://doi.org/10.1029/2011GL046893>.
- Hu, Z., S. Yang, and R. Wu, 2003: Long-term climate variations in China and global warming signals. *J. Geophys. Res.*, **108**, 4614, <https://doi.org/10.1029/2003JD003651>.
- Jiang, J., T. Zhou, B. Wu, and L. Zou, 2019: Ocean–atmosphere interaction processes over the northwestern Pacific simulated by an ocean data assimilation system based on a coupled model FGOALS-s2 (in Chinese). *Chin. J. Atmos. Sci.*, **43**, 467–482, <https://doi.org/10.3878/j.issn.1006-9895.1806.18146>.
- Kendall, M., 1955: *Rank Correlation Methods*. Oxford University Press, 196 pp.
- Kiehl, J. T., J. J. Hack, G. B. Bonan, B. A. Boville, D. L. Williamson, and P. J. Rasch, 1998: The National Center for Atmospheric Research Community Climate Model: CCM3. *J. Climate*, **11**, 1131–1149, [https://doi.org/10.1175/1520-0442\(1998\)011<1131:TNCFAR>2.0.CO;2](https://doi.org/10.1175/1520-0442(1998)011<1131:TNCFAR>2.0.CO;2).
- Kitoh, A., H. Endo, K. Krishna Kumar, I. F. A. Cavalcanti, P. Goswami, and T. Zhou, 2013: Monsoons in a changing world: A regional perspective in a global context. *J. Geophys. Res.*, **118**, 3053–3065, <https://doi.org/10.1002/JGRD.50258>.
- Kripalani, R. H., A. Kulkarni, S. S. Sabade, and M. L. Khandekar, 2003: Indian monsoon variability in a global warming scenario. *Nat. Hazards*, **29**, 189–206, <https://doi.org/10.1023/A:1023695326825>.
- Krishnamurthy, L., and V. Krishnamurthy, 2014: Influence of PDO on South Asian summer monsoon and monsoon–ENSO relation. *Climate Dyn.*, **42**, 2397–2410, <https://doi.org/10.1007/s00382-013-1856-z>.
- , and —, 2016: Teleconnections of Indian monsoon rainfall with AMO and Atlantic tripole. *Climate Dyn.*, **46**, 2269–2285, <https://doi.org/10.1007/s00382-015-2701-3>.
- , and —, 2017: Indian monsoon's relation with the decadal part of PDO in observations and NCAR CCSM4. *Int. J. Climatol.*, **37**, 1824–1833, <https://doi.org/10.1002/joc.4815>.
- Krishnamurthy, V., and B. N. Goswami, 2000: Indian monsoon–ENSO relationship on interdecadal timescale. *J. Climate*, **13**, 579–595, [https://doi.org/10.1175/1520-0442\(2000\)013<0579:IMEROI>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<0579:IMEROI>2.0.CO;2).
- , and B. P. Kirtman, 2003: Variability of the Indian Ocean: Relation to monsoon and ENSO. *Quart. J. Roy. Meteor. Soc.*, **129**, 1623–1646, <https://doi.org/10.1256/qj.01.166>.
- Kucharski, F., F. Molteni, and J. H. Yoo, 2006: SST forcing of decadal Indian monsoon rainfall variability. *Geophys. Res. Lett.*, **33**, L03709, <https://doi.org/10.1029/2005GL025371>.
- , and Coauthors, 2009: The CLIVAR C20C project: Skill of simulating Indian monsoon rainfall on interannual to decadal timescales. Does GHG forcing play a role? *Climate Dyn.*, **33**, 615–627, <https://doi.org/10.1007/s00382-008-0462-y>.
- Kumar, S., V. Merwade, J. L. Kinter, and D. Niyogi, 2013: Evaluation of temperature and precipitation trends and long-term persistence in CMIP5 twentieth-century climate simulations. *J. Climate*, **26**, 4168–4185, <https://doi.org/10.1175/JCLI-D-12-00259.1>.
- Li, H., A. Dai, T. Zhou, and J. Lu, 2010: Responses of East Asian summer monsoon to historical SST and atmospheric forcing during 1950–2000. *Climate Dyn.*, **34**, 501–514, <https://doi.org/10.1007/s00382-008-0482-7>.
- Li, Y., Y. Ding, and W. Li, 2017: Interdecadal variability of the Afro-Asian summer monsoon system. *Adv. Atmos. Sci.*, **34**, 833–846, <https://doi.org/10.1007/s00376-017-6247-7>.
- Li, Z., F. Niu, J. Fan, Y. Liu, D. Rosenfeld, and Y. Ding, 2011: Long-term impacts of aerosols on the vertical development of clouds and precipitation. *Nat. Geosci.*, **4**, 888–894, <https://doi.org/10.1038/ngeo1313>.
- Lin, R., T. Zhou, and Y. Qian, 2014: Evaluation of global monsoon precipitation changes based on five reanalysis datasets. *J. Climate*, **27**, 1271–1289, <https://doi.org/10.1175/JCLI-D-13-00215.1>.
- Liu, Y., and J. C. H. Chiang, 2012: Coordinated abrupt weakening of the Eurasian and North African monsoons in the 1960s and links to extratropical North Atlantic cooling. *J. Climate*, **25**, 3532–3548, <https://doi.org/10.1175/JCLI-D-11-00219.1>.
- Matsumoto, T., 1966: Quasi-geostrophic motions in the equatorial area. *J. Meteor. Soc. Japan*, **44**, 25–43, https://doi.org/10.2151/jmsj1965.44.1_25.
- Meehl, G. A., and A. Hu, 2006: Megadroughts in the Indian monsoon region and southwest North America and a mechanism for associated multidecadal Pacific sea surface temperature anomalies. *J. Climate*, **19**, 1605–1623, <https://doi.org/10.1175/JCLI3675.1>.
- , and J. M. Arblaster, 2011: Decadal variability of Asian–Australian monsoon–ENSO–TBO relationships. *J. Climate*, **24**, 4925–4940, <https://doi.org/10.1175/2011JCLI4015.1>.
- Mohino, E., S. Janicot, and J. Bader, 2011: Sahel rainfall and decadal to multi-decadal sea surface temperature variability. *Climate Dyn.*, **37**, 419–440, <https://doi.org/10.1007/s00382-010-0867-2>.
- Polson, D., M. Bollasina, G. C. Hegerl, and L. J. Wilcox, 2014: Decreased monsoon precipitation in the Northern Hemisphere due to anthropogenic aerosols. *Geophys. Res. Lett.*, **41**, 6023–6029, <https://doi.org/10.1002/2014GL060811>.
- Pope, V. D., M. L. Gallani, P. R. Rowntree, and R. A. Stratton, 2000: The impact of new physical parametrizations in the Hadley Centre climate model: HadAM3. *Climate Dyn.*, **16**, 123–146, <https://doi.org/10.1007/s003820050009>.
- Power, S., T. Casey, C. Folland, A. Colman, and V. Mehta, 1999: Inter-decadal modulation of the impact of ENSO on Australia. *Climate Dyn.*, **15**, 319–324, <https://doi.org/10.1007/s003820050284>.
- Qian, C., and T. Zhou, 2014: Multidecadal variability of North China aridity and its relationship to PDO during 1900–2010. *J. Climate*, **27**, 1210–1222, <https://doi.org/10.1175/JCLI-D-13-00235.1>.
- Ramanathan, V., P. J. Crutzen, J. T. Kiehl, and D. Rosenfeld, 2001: Aerosols, climate, and the hydrological cycle. *Science*, **294**, 2119–2124, <https://doi.org/10.1126/science.1064034>.
- Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan, 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.*, **108**, 4407, <https://doi.org/10.1029/2002JD002670>.
- Scaife, A. A., and Coauthors, 2009: The CLIVAR C20C Project: Selected 20th century climate events. *Climate Dyn.*, **33**, 603–614, <https://doi.org/10.1007/s00382-008-0451-1>.
- Schmidt, G. A., and Coauthors, 2014: Configuration and assessment of the GISS ModelE2 contributions to the CMIP5 archive. *J. Adv. Model. Earth Syst.*, **6**, 141–184, <https://doi.org/10.1002/2013MS000265>.
- Schneider, T., T. Bischoff, and G. H. Haug, 2014: Migrations and dynamics of the intertropical convergence zone. *Nature*, **513**, 45–53, <https://doi.org/10.1038/nature13636>.
- Schneider, U., A. Becker, P. Finger, A. Meyer-Christoffer, B. Rudolf, and M. Ziese, 2014: GPCC full data monthly product version 7.0 at 1.0°: Monthly land-surface precipitation from rain-gauges built on GTS-based and historic data. Accessed

- 28 November 2018, https://doi.org/10.5676/DWD_GPCC/FD_M_V7_100.
- Schubert, S., and Coauthors, 2009: A U.S. CLIVAR project to assess and compare the responses of global climate models to drought-related SST forcing patterns: Overview and results. *J. Climate*, **22**, 5251–5272, <https://doi.org/10.1175/2009JCLI3060.1>.
- , and Coauthors, 2016: Global meteorological drought: A synthesis of current understanding with a focus on SST drivers of precipitation deficits. *J. Climate*, **29**, 3989–4019, <https://doi.org/10.1175/JCLI-D-15-0452.1>.
- Seager, R., N. Naik, and G. A. Vecchi, 2010: Thermodynamic and dynamic mechanisms for large-scale changes in the hydrological cycle in response to global warming. *J. Climate*, **23**, 4651–4668, <https://doi.org/10.1175/2010JCLI3655.1>.
- Sheffield, J., E. F. Wood, and M. L. Roderick, 2012: Little change in global drought over the past 60 years. *Nature*, **491**, 435–438, <https://doi.org/10.1038/nature11575>.
- Song, F., T. Zhou, and Y. Qian, 2014: Responses of East Asian summer monsoon to natural and anthropogenic forcings in the 17 latest CMIP5 models. *Geophys. Res. Lett.*, **41**, 596–603, <https://doi.org/10.1002/2013GL058705>.
- Sun, Q., C. Miao, Q. Duan, H. Ashouri, S. Sorooshian, and K. L. Hsu, 2018: A review of global precipitation data sets: Data sources, estimation, and intercomparisons. *Rev. Geophys.*, **56**, 79–107, <https://doi.org/10.1002/2017RG000574>.
- Sutton, R. T., and D. L. R. Hodson, 2005: Atlantic Ocean forcing of North American and European summer climate. *Science*, **309**, 115–118, <https://doi.org/10.1126/science.1109496>.
- Trenberth, K. E., D. P. Stepaniak, and J. M. Caron, 2000: The global monsoon as seen through the divergent atmospheric circulation. *J. Climate*, **13**, 3969–3993, [https://doi.org/10.1175/1520-0442\(2000\)013<3969:TGMASST>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<3969:TGMASST>2.0.CO;2).
- , A. Dai, G. Van Der Schrier, P. D. Jones, J. Barichivich, K. R. Briffa, and J. Sheffield, 2014: Global warming and changes in drought. *Nat. Climate Change*, **4**, 17–22, <https://doi.org/10.1038/nclimate2067>.
- van Oldenborgh, G. J., F. J. Doblas-Reyes, B. Wouters, and W. Hazeleger, 2012: Decadal prediction skill in a multi-model ensemble. *Climate Dyn.*, **38**, 1263–1280, <https://doi.org/10.1007/s00382-012-1313-4>.
- Wang, B., and Q. Ding, 2006: Changes in global monsoon precipitation over the past 56 years. *Geophys. Res. Lett.*, **33**, L06711, <https://doi.org/10.1029/2005GL025347>.
- , and —, 2008: Global monsoon: Dominant mode of annual variation in the tropics. *Dyn. Atmos. Oceans*, **44**, 165–183, <https://doi.org/10.1016/j.dynatmoce.2007.05.002>.
- , —, X. Fu, I. S. Kang, K. Jin, J. Shukla, and F. Doblas-Reyes, 2005: Fundamental challenge in simulation and prediction of summer monsoon rainfall. *Geophys. Res. Lett.*, **32**, L15711, <https://doi.org/10.1029/2005GL022734>.
- , J. Liu, H. J. Kim, P. J. Webster, and S. Y. Yim, 2012: Recent change of the global monsoon precipitation (1979–2008). *Climate Dyn.*, **39**, 1123–1135, <https://doi.org/10.1007/s00382-011-1266-z>.
- , —, —, —, —, and B. Xiang, 2013: Northern Hemisphere summer monsoon intensified by mega-El Niño/southern oscillation and Atlantic multidecadal oscillation. *Proc. Natl. Acad. Sci. USA*, **110**, 5347–5352, <https://doi.org/10.1073/pnas.1219405110>.
- , and Coauthors, 2018: Toward predicting changes in the land monsoon rainfall a decade in advance. *J. Climate*, **31**, 2699–2714, <https://doi.org/10.1175/JCLI-D-17-0521.1>.
- Wang, P. X., B. Wang, H. Cheng, J. Fasullo, Z. T. Guo, T. Kiefer, and Z. Y. Liu, 2017: The global monsoon across time scales: Mechanisms and outstanding issues. *Earth Sci. Rev.*, **174**, 84–121, <https://doi.org/10.1016/j.earscirev.2017.07.006>.
- Wilks, D. S., 1995: *Statistical Methods in the Atmospheric Sciences: An Introduction*. Academic Press, 467 pp.
- Willmott, C. J., and K. Matsuura, 2000: Terrestrial air temperature and precipitation: Monthly and annual climatologies. Centre for Climate Research, Department of Geography, University of Delaware, accessed 4 December 2017, http://climate.geog.udel.edu/~climate/html_pages/Global2_Clim/README.global2_clim.html.
- Xie, P., and P. A. Arkin, 1997: Global precipitation: A 17-year monthly analysis based on gauge observations, satellite estimates, and numerical model outputs. *Bull. Amer. Meteor. Soc.*, **78**, 2539–2558, [https://doi.org/10.1175/1520-0477\(1997\)078<2539:GPAYMA>2.0.CO;2](https://doi.org/10.1175/1520-0477(1997)078<2539:GPAYMA>2.0.CO;2).
- Yang, Y., Y. Du, Y. Wu, G. Huang, and Y. Zhang, 2015: The interannual variations of summer precipitation in the northern Indian Ocean associated with ENSO. *Atmos. Oceanic Sci. Lett.*, **5**, 301–305, <https://doi.org/10.1080/16742834.2012.11447014>.
- Yin, X., A. Gruber, and P. Arkin, 2004: Comparison of the GPCP and CMAP merged gauge–satellite monthly precipitation products for the period 1979–2001. *J. Hydrometeor.*, **5**, 1207–1222, <https://doi.org/10.1175/JHM-392.1>.
- Yu, R., and T. Zhou, 2007: Seasonality and three-dimensional structure of the interdecadal change in East Asian monsoon. *J. Climate*, **20**, 5344–5355, <https://doi.org/10.1175/2007JCLI1559.1>.
- , B. Wang, and T. Zhou, 2004: Tropospheric cooling and summer monsoon weakening trend over East Asia. *Geophys. Res. Lett.*, **31**, L22212, <https://doi.org/10.1029/2004GL021270>.
- Zhang, L., and T. Zhou, 2011: An assessment of monsoon precipitation changes during 1901–2001. *Climate Dyn.*, **37**, 279–296, <https://doi.org/10.1007/s00382-011-0993-5>.
- Zhang, R., and T. L. Delworth, 2006: Impact of Atlantic multidecadal oscillations on India/Sahel rainfall and Atlantic hurricanes. *Geophys. Res. Lett.*, **33**, L17712, <https://doi.org/10.1029/2006GL026267>.
- Zhou, T., L. Zhang, and H. Li, 2008a: Changes in global land monsoon area and total rainfall accumulation over the last half century. *Geophys. Res. Lett.*, **35**, L16707, <https://doi.org/10.1029/2008GL034881>.
- , R. Yu, H. Li, and B. Wang, 2008b: Ocean forcing to changes in global monsoon precipitation over the recent half-century. *J. Climate*, **21**, 3833–3852, <https://doi.org/10.1175/2008JCLI2067.1>.
- , and Coauthors, 2009a: The CLIVAR C20C Project: Which components of the Asian–Australian monsoon circulation variations are forced and reproducible? *Climate Dyn.*, **33**, 1051–1068, <https://doi.org/10.1007/s00382-008-0501-8>.
- , B. Wu, and B. Wang, 2009b: How well do atmospheric general circulation models capture the leading modes of the interannual variability of the Asian–Australian monsoon? *J. Climate*, **22**, 1159–1173, <https://doi.org/10.1175/2008JCLI2245.1>.
- , D. Gong, J. Li, and B. Li, 2009c: Detecting and understanding the multi-decadal variability of the East Asian summer monsoon—Recent progress and state of affairs. *Meteor. Z.*, **18**, 455–467, <https://doi.org/10.1127/0941-2948/2009/0396>.
- , and Coauthors, 2016: GMMIP (v1.0) contribution to CMIP6: Global Monsoons Model Intercomparison Project. *Geosci. Model Dev.*, **9**, 3589–3604, <https://doi.org/10.5194/gmd-9-3589-2016>.

- Zhu, Y., H. Wang, W. Zhou, and J. Ma, 2011: Recent changes in the summer precipitation pattern in East China and the background circulation. *Climate Dyn.*, **36**, 1463–1473, <https://doi.org/10.1007/s00382-010-0852-9>.
- , —, J. Ma, T. Wang, and J. Sun, 2015: Contribution of the phase transition of Pacific decadal oscillation to the late 1990s' shift in East China summer rainfall. *J. Geophys. Res.*, **120**, 8817–8827, <https://doi.org/10.1002/2015JD023545>.
- Zou, L., and T. Zhou, 2013: Can a regional ocean–atmosphere coupled model improve the simulation of the interannual variability of the western North Pacific summer monsoon? *J. Climate*, **26**, 2353–2367, <https://doi.org/10.1175/JCLI-D-11-00722.1>.
- , and —, 2016: A regional ocean–atmosphere coupled model developed for CORDEX East Asia: Assessment of Asian summer monsoon simulation. *Climate Dyn.*, **47**, 3627–3640, <https://doi.org/10.1007/s00382-016-3032-8>.
- , —, and D. Peng, 2016: Dynamical downscaling of historical climate over CORDEX East Asia domain: A comparison of regional ocean atmosphere coupled model to standalone RCM simulations. *J. Geophys. Res. Atmos.*, **121**, 1442–1458, <https://doi.org/10.1002/2015JD023912>.
- , D. Li, T. Zhou, and B. Wu, 2018: Summer rainfall–SST relationships in the western North Pacific simulated by the FGOALS model with ocean assimilation (in Chinese). *Climatic Environ. Res.*, **23**, 139–149.