Weakened Impact of the Developing El Niño on Tropical Indian Ocean Climate Variability under Global Warming

CHAO HE

Institute for Environmental and Climate Research, Jinan University, Guangzhou, and Key Laboratory of Meteorological Disaster of Ministry of Education, Nanjing University of Information Science and Technology, Nanjing, China

YUHAO WANG

Key Laboratory of Meteorological Disaster of Ministry of Education, Nanjing University of Information Science and Technology, Nanjing, China

TIM LI

Key Laboratory of Meteorological Disaster of Ministry of Education, Nanjing University of Information Science and Technology, Nanjing, China, and International Pacific Research Center, and Department of Atmospheric Sciences, University of Hawai'i at Mānoa, Honolulu, Hawaii

(Manuscript received 28 February 2019, in final form 22 June 2019)

ABSTRACT

El Niño induces an anomalous easterly wind along the equator and a pair of anomalous anticyclones straddling the equator over the tropical Indian Ocean (TIO) during the autumn of its developing phase. Based on 30 coupled models participating in CMIP5, these atmospheric circulation anomalies over TIO are substantially weakened by about 12%–13% K^{-1} under global warming scenarios, associated with a weakened zonal gradient of the sea surface temperature (SST) anomaly. The mechanism for the response is investigated based on a hierarchy of model experiments. Based on stand-alone atmospheric model experiments under uniform and patterned mean-state SST warming, the atmospheric circulation anomaly over TIO during the autumn of the developing El Niño is also substantially weakened by about 8% K^{-1} even if the interannual variability of SST remains exactly unchanged, suggesting that the primary cause resides in the atmosphere rather than the SST anomaly. The tropospheric static stability is robustly enhanced under global warming, and experiments performed by a linear baroclinic model show that a much weaker atmospheric circulation anomaly over TIO is stimulated by an unchanged diabatic heating anomaly under a more stable atmosphere. The weakened atmospheric circulation anomaly due to enhanced static stability weakens the zonal gradient of the SST anomaly within TIO through local air–sea interaction, and it acts to further weaken the atmospheric circulation anomaly. The enhanced static stability of the troposphere is probably the primary cause and the air–sea interaction within TIO is a secondary cause for the weakened impact of the developing El Niño on atmospheric circulation variability over TIO.

1. Introduction

El Niño–Southern Oscillation (ENSO) is the dominant signal of interannual climate variability in the tropics, and it has a substantial impact on the climate variability over the tropical Indian Ocean (TIO)

through an atmospheric bridge (Lau and Nath 2003; Yang et al. 2018). During the El Niño developing phase, atmospheric circulation over TIO is characterized by an easterly wind anomaly along the equator, associated with a pair of anomalous anticyclones straddling the equator over the northern and the southern TIO (Fig. 1a). These atmospheric circulation anomalies are stimulated by the negative latent heating anomaly around the Maritime Continent (MC), reminiscent of a Gill-type response (Gill 1980; Lau and Nath 2003; Li et al. 2003). The anomalous atmospheric circulation during El Niño's developing phase is responsible for

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-19-0165.s1>.

Corresponding author: Dr. Tim Li, timli@hawaii.edu

FIG. 1. The atmospheric anomalies regressed onto the Niño-3.4 index in SON for the (top) observation, (middle) Historical experiment by CGCMs, and (bottom) AMIP experiment by AGCMs. Shown are the (a)–(c) 850-hPa wind and H_e anomalies, (d)–(f) 850-hPa ζ anomaly, and (g)–(i) precipitation anomaly. The regressed anomalies significant at the 95% confidence level are stippled.

deficient rainfall over the South Asian monsoon area (Kumar et al. 1999; Xavier et al. 2007; Cash et al. 2017), and vice versa during the developing phase of La Niña.

El Niño has an impact on TIO climate variability not only in the atmosphere but also the atmosphere–ocean coupled system. The Indian Ocean dipole (IOD) is the dominant mode of interannual variability of sea surface temperature (SST) over TIO in autumn, and a positive (negative) IOD event is characterized by a negative (positive) SST anomaly over the eastern TIO and positive (negative) SST anomaly over the western TIO (Saji et al. 1999; Yamagata et al. 2004). The IOD is impacted by the atmospheric circulation anomaly stimulated by ENSO (Li et al. 2003; Wang and Wang 2014; Fan et al. 2017; Liu et al. 2017; Hu et al. 2018), and an El Niño (La Niña) event may lead to a positive (negative) phase of IOD event primarily through the easterly (westerly) wind anomaly over the equatorial Indian Ocean (Li et al. 2003; Feng and Duan 2018). The formation of the IOD event further enhances the atmospheric circulation anomaly over TIO (Li et al. 2003), and may further modulate the interannual climate variability over the Asian monsoon region (Kripalani et al.

2010; Ummenhofer et al. 2011; Cherchi and Navarra 2013; Xu et al. 2016) and tropical Pacific Ocean (Kug et al. 2006; Luo et al. 2010; Izumo et al. 2014).

The association of climate variability over TIO with ENSO has experienced decadal changes. Observational records show that the impact of the developing El Niño on South Asian summer monsoon rainfall is weakened in recent decades (Kumar et al. 1999; Xavier et al. 2007), and the coupling between ENSO and IOD variability is also weakened (Ham et al. 2017). All this evidence suggests a possibly weakened impact of the developing El Niño on the climate variability over TIO, but it is unclear whether such decadal climate change results from global warming or the internal variability of the climate system, since observed decadal change is affected by multiple factors, especially the internal variability of the climate system (Deser et al. 2012; Hui and Zheng 2018; Maher et al. 2018; Wang et al. 2019).

To understand the effect of global warming due to increasing greenhouse gases (GHG), ensemble simulations by coupled general circulation models (CGCMs) are needed to suppress the stochastic oscillation generated by internal variability (Deser et al. 2012). Based on

the ensemble simulations by CGCMs under increased GHG concentration, the response of ENSO to global warming is subject to great uncertainty (Collins et al. 2010; Chen et al. 2015, 2017; Maher et al. 2018). Even if El Niño itself remains unchanged, El Niño can stimulate a stronger precipitation anomaly over the tropical Pacific Ocean during its mature phase (Power et al. 2013; Cai et al. 2014) but a weaker atmospheric circulation anomaly over the western North Pacific (WNP) during its decaying phase (Jiang et al. 2018; He et al. 2019). However, it is still unclear how global warming modulates the El Niño-induced climate variability over TIO during El Niño developing phase, and it motivates us to perform this study.

This study focuses on the response of atmospheric circulation anomalies over TIO during the autumn of the El Niño developing phase, and the remainder of this paper is organized as follows. The model, data, and method are introduced in section 2, and a brief model evaluation is performed in section 3. The response of the El Niño-induced atmospheric circulation anomaly over TIO and the mechanism are addressed in section 4, based on a hierarchy of model simulations, and the conclusions are summarized in section 5.

2. Model, data, and method

a. CMIP5 models

Thirty CGCMs participating in phase 5 of the Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012) are adopted for analysis. The representative concentration pathway (RCP) experiments performed by these CGCMs are compared with their Historical experiments, to extract the GHG-forced response. The Historical experiment is driven by observed concentrations of external forcing (e.g., GHGs, aerosols) from 1850 to 2005, and the period of 1950–99 is adopted in this study and referred to as 20C. The RCP8.5 experiment is driven by a business-as-usual pathway of emission toward a radiative forcing of 8.5 W m^{-2} until 2100, and the RCP4.5 experiment is driven by a mitigation pathway toward a radiative forcing of 4.5 W m^{-2} until 2100 (Van Vuuren et al. 2011). The 2050–99 epochs of RCP8.5 and RCP4.5 experiments are adopted in this study and referred to as 21C. Only the first realization of each model and each experiment is analyzed to equally weigh each model. Most of the results are based on the RCP8.5 scenario, and the RCP4.5 scenario is adopted to examine the possible dependence of the results on the scenario.

To investigate whether the change in the El Niño-related SST anomaly (SSTA) plays a key role for the

CGCM-projected change of the atmospheric circulation anomaly over TIO, the AMIP4K and AMIPFuture experiments based on 11 atmospheric general circulation models (AGCMs) participating CMIP5 (CanAM4, HadGEM2-A, and the nine boldface models listed in Table S1 in the online supplemental material) are compared with their AMIP experiments. The AMIP experiment is forced by observed global monthly SST from 1979 to 2008. The SSTs used as the lower-boundary condition of AGCMs in the AMIP4K and AMIPFuture experiments are obtained by adding a time-invariant global increase of SST to observed monthly SST, where a spatially uniform warming of 4 K is prescribed in the AMIP4K experiment and a patterned warming is prescribed in the AMIPFuture experiment (Taylor et al. 2012; Zhou et al. 2014). The interannual variability of SST is exactly the same in the AMIP, AMIP4K, and AMIPFuture experiments after removing the SST climatology. A comparison of the AMIP4K (or AMIPFuture) experiment with the AMIP experiment tells about the response of the interannual climate variability to global warming if the El Niño-related SSTA remains unchanged, and a comparison between the AMIPFuture and AMIP4K experiments tells whether the change in the interannual climate variability depends on the warming pattern of mean-state SST. The AMIP experiment is referred to as 20C and the AMIP4K and AMIPFuture experiments are referred to as 21C in the discussion.

b. Observational data

To evaluate the models against the observation, the following monthly observational datasets (including reanalysis data) are adopted. 1) The atmospheric variables are from the NCEP–NCAR reanalysis (Kalnay et al. 1996), including geopotential height, horizontal wind, and vertical velocity. Relative vorticity ζ is calculated from the horizontal wind vector, and eddy geopotential height (H_e) is calculated by removing the simultaneous zonal-mean geopotential height to eliminate its systematic increase under global warming (He et al. 2018). The three-dimensional diabatic heating is calculated following the thermodynamic equation based on NCEP–NCAR reanalysis data (Yanai and Tomita 1998). 2) The monthly SST is derived from the Extended Reconstructed SST, version 5 (Huang et al. 2017). 3) The gridded precipitation spanning from 1979 to 2008 is derived from the Global Precipitation Climatology Project (GPCP), version 2.3 (Adler et al. 2003). The model data are bilinearly interpolated onto a $2.5^\circ \times 2.5^\circ$ horizontal grid identical to the NCEP–NCAR reanalysis, and we focus on boreal autumn [September–November (SON)] throughout this study.

c. Method

Based on both CGCM and AGCM experiments, an 8-yr high-pass Fourier filter is applied to the original time series of all variables before analyses, to obtain the interannual variability component. As there is great uncertainty in the change of SSTA of ENSO (Collins et al. 2010; Chen et al. 2015, 2017; Maher et al. 2018), this study only focuses on the sensitivity of the atmospheric circulation anomaly to ENSO but not the change of ENSO. Linear regression of the seasonal-mean anomaly onto the contemporary Niño-3.4 index (regional averaged SSTA over 5°S–5°N, 170°–120°W subject to an 8-yr high-pass filter, not standardized) is used to measure the El Niño-related climate variability, and the case for La Niña is vice versa and not particularly discussed. The regressed anomalies at the interannual time scale are calculated for both 20C and the 21C, and the difference in the regression slopes between 21C and 20C is scaled by the amplitude of tropical-mean SST warming (2.49 K for RCP8.5 and 1.51 K for RCP4.5 experiment relative to the Historical experiment; 4.0 K for AMIP4K and 4.4 K for AMIPFuture experiment relative to the AMIP experiment), to quantify the change in the interannual variability per degree of tropical-mean SST warming. The linear regression is performed for each model, respectively, and the multimodel median (MMM) of the change is adopted to suppress the internal variability and stochastic model error, and the MMM is regarded as the response to global warming. MMM is superior to the multimodel mean value since it is not sensitive to outliers (Gleckler et al. 2008). The significance of the response is determined by whether more than 70% of the individual models agree on the sign of the change, as an intermodel consensus of 68% is equivalent to the 95% confidence level of the Student's *t* test (Power et al. 2012).

d. Linear baroclinic model

The CGCMs and AGCMs are designed to realistically simulating Earth's climate by considering the known physical processes as many as possible (e.g., radiation, cloud, and boundary layer processes). It helps to simulate the overall effect of all the considered processes on climate change by using the complex climate models, but it is difficult to understand the essential mechanism responsible for the climate change. In contrast to the complex AGCMs, the linear baroclinic model (LBM) is a simple linear anomaly model developed by Watanabe and Kimoto (2000). The key difference between LBM and AGCM is that there are no physical processes in LBM and the dynamic core is simplified and linearized. The LBM can be directly used to examine the

response of atmospheric circulation anomaly stimulated by a prescribed diabatic heating anomaly under a certain mean state. In this study, the impact of the mean state on the atmospheric circulation anomalies stimulated by a diabatic heating anomaly are investigated by using LBM with a T42 horizontal grid and 20 vertical levels.

3. Observed and simulated TIO climate variability during El Niño developing in autumn

The atmospheric circulation anomalies during the autumn of the El Niño developing phase simulated by the MMM of CGCM and AGCM are evaluated against the observation in Fig. 1, by regressing the anomalies of atmospheric variables onto the contemporary Niño-3.4 index in SON for the observation, Historical, and AMIP experiments, respectively. During the autumn of the El Niño developing phase, an anomalous easterly wind prevails from MC to equatorial Indian Ocean, associated with a pair of anomalous anticyclones on the northern and southern sides of the equator over TIO (Fig. 1a). This large-scale pattern of the wind anomaly is captured by both CGCM and AGCM simulations (Figs. 1b,c). The easterly wind anomaly along the equatorial Indian Ocean is essential for the development of the positive IOD event, which usually appears during the El Niño developing phase (Li et al. 2003; Yuan and Li 2008; Feng and Duan 2018). The anomalous anticyclones straddling the equator over TIO are characterized by a negative relative vorticity ζ anomaly over the northern TIO and positive ζ anomaly over the southern TIO (Fig. 1d). Similar to the observation, the anomalous ζ field during the autumn of the El Niño developing phase is also characterized by a significant negative (positive) anomaly on the northern (southern) side of the equator over TIO (Figs. 1e,f).

The most prominent feature of the precipitation anomaly during the autumn of the developing El Niño is the negative precipitation anomaly around MC, which is a robust feature in the observation and CGCM and AGCM simulations (Figs. 1g–i). The negative latent heating anomaly associated with precipitation anomaly around MC is responsible for the easterly wind anomaly along equatorial Indian Ocean and the anomalous anticyclone pair over TIO (Gill 1980), since the tropical circulation anomaly is directly driven by the latent heating anomaly associated with deep convection. The negative precipitation anomaly over MC is directly forced by the anomalous subsidence associated with the anomalous Walker circulation, and the anomalous Walker circulation is driven by the heating anomaly over the equatorial Pacific (Lau and Nath 2003; Yang et al. 2018). The similar pattern of the precipitation

FIG. 2. Scatter diagram for the regressed anomalies of regional averaged quantities onto the Niño-3.4 index in SON simulated by the individual (a),(b) CGCMs and (c),(d) AGCMs. (left) The U index anomaly over the equatorial Indian Ocean as a function of the P index anomaly over MC. (right) The ζ index anomaly over TIO as a function of the P index anomaly over MC. The blue dot, red circle, and green star indicate the individual models, the MMM, and the observation, respectively. The intermodel correlation coefficients among the individual models are marked in the upper-right corner of each panel, and the correlation coefficients significant at the 95% confidence level are marked with an asterisk.

anomaly between model and observation indicates that the mechanism for the formation of the atmospheric circulation anomalies over TIO is captured by the models. Overall, the El Niño-induced atmospheric anomalies over TIO and the negative heating anomaly over MC are captured by the MMM of CGCMs and AGCMs.

A quantitative evaluation on the MMM and individual models against the observation is performed, in terms of the regional averaged anomalies. The regional averaged precipitation anomaly over MC (10°S – 10°N , 80° – 120°E) is defined as the P index, the regional

averaged 850-hPa zonal wind anomaly over the equatorial Indian Ocean (8°S – 8°N , 60° – 100°E) is defined as the U index, and the difference of the ζ anomaly between northern TIO (2° – 15°N , 60° – 100°E) and southern TIO (2° – 15°S , 60° – 100°E) is defined as the ζ index. These three indices are regressed onto the Niño-3.4 index to quantitatively measure the El Niño-induced climate variability, and the scatterplots for the individual models and the MMM are shown in Fig. 2, in comparison with the observation.

The negative precipitation anomaly over MC, the easterly wind anomaly over the equatorial Indian Ocean,

and the anticyclone anomalies over the off-equatorial TIO are captured by these three indices in all the 30 CGCMs (Figs. 2a,b) and 11 AGCMs (Figs. 2c,d). There is no obvious systematic bias in the CGCM, but the anomalies in the AGCMs are weaker than the observation. The MMM of the atmospheric circulation anomalies in the CGCMs are very close to the observation (Figs. 2a,b), although the anomalous easterly wind along the equator in the MMM is weaker than the observation over the eastern Indian Ocean but stronger than the observation over the western Indian Ocean (see Fig. S1). The amplitudes of the atmospheric circulation anomalies are systematically underestimated in the AGCMs (Figs. 2c,d), consistent with the spatial pattern of the bias for the MMM (see Fig. S1). The underestimation of the atmospheric circulation anomalies over TIO are also consistent with the underestimated negative precipitation anomaly over MC in AGCMs (Figs. 2c,d).

In the CGCMs, the anomalous precipitation over MC has a high intermodel correlation of 0.83 with the atmospheric circulation anomalies over TIO (Figs. 2a,b). The models with a stronger negative precipitation anomaly over MC are characterized by a stronger easterly wind anomaly over the equatorial Indian Ocean and stronger anomalous anticyclones over the northern and southern TIO, suggesting that the latent heating anomaly over MC well explains the intermodel spread in the atmospheric circulation anomalies over TIO. But the precipitation anomaly over MC cannot well explain the intermodel spread of the atmospheric circulation anomalies over TIO in the AGCMs, as the positive intermodel correlation coefficients are insignificant at the 95% confidence level (Figs. 2c,d).

Overall, the El Niño-induced atmospheric circulation anomalies over TIO resemble a Gill-type response to the negative latent heating anomaly over MC (Gill 1980), in both the observation and the CMIP5 models. This is consistent with the mechanism for the impact of El Niño on Indian Ocean climate variability during its developing phase proposed by previous studies (Lau and Nath 2003; Li et al. 2003; Feng and Duan 2018), giving us confidence to further investigate the response of El Niño-induced TIO climate variability to global warming.

4. Responses of El Niño-induced TIO variability to global warming

a. Projected changes by CGCM

Figure 3 shows the projected changes in the regressed atmospheric circulation anomalies onto the Niño-3.4

index under RCP8.5, scaled by tropical-mean SST warming. The change in the low-level wind anomaly over TIO is characterized by an anomalous westerly wind along the equator and two anomalous cyclones on both sides of the equator, consistent with the change in H_e anomaly (Fig. 3a), and the pattern of change is almost opposite to the atmospheric circulation anomaly during 20C (see Fig. 1). The change in the El Niño-induced ζ anomaly over TIO is characterized by positive values over the northern TIO and negative values over the southern TIO off the equator (Fig. 3b), which is also opposite to the El Niño-induced ζ anomaly in 20C. The above changes suggest that the TIO circulation anomaly associated with a developing El Niño is weaker under a warmer climate. A similar pattern of response is obtained under the RCP4.5 scenario (Fig. S2).

The change in the precipitation anomaly responsible for latent heating is examined in Fig. 3c. The negative precipitation anomaly off the southwest coast of Sumatra is substantially weakened, but the change of precipitation anomaly over other parts of MC is ambiguous and insignificant, which remains to be quantitatively evaluated. The change in the precipitation anomaly over TIO is characterized by a weakened zonal gradient of precipitation anomaly across the TIO basin, consistent with the weakened easterly wind anomaly over the equatorial Indian Ocean (Saji et al. 1999; Li et al. 2003). As vertical moisture advection induced by anomalous vertical motion is the most important contributor to the tropical precipitation anomaly (Ni and Hsu 2018; He and Li 2019), the change in the vertical velocity anomaly at 500 hPa during the El Niño developing in autumn is shown in Fig. 3d. Compared with 20C (contours in Fig. 3d), the change in the vertical velocity anomaly is characterized by weakened descending anomaly off the southwest coast of Sumatra and weakened ascending anomaly over the equatorial western Indian Ocean (shading in Fig. 3d). Based on He and Li (2019), the precipitation anomaly is reconstructed by the product of mean-state specific humidity at 925 hPa ($\bar{q}$) and vertical velocity anomaly at 500 hPa (ω'), that is, $P' \approx -\bar{q}\omega'/g$. The projected change in the anomaly of $-\bar{q}\omega'/g$ during the autumn of the developing El Niño (Fig. 3e) resembles Fig. 3c, suggesting the dominant contribution of vertical moisture advection to the change in the precipitation anomaly over TIO.

As there is an active air-sea coupling within TIO (Saji et al. 1999; Li et al. 2003; Yamagata et al. 2004), the projected change in SSTA during El Niño developing in autumn is examined in Fig. 3f. Compared with 20C (contours in Fig. 3f), the negative SSTA around MC during El Niño developing in autumn is weakened in

FIG. 3. The MMM-projected changes in regressed atmospheric and oceanic anomalies onto the Niño-3.4 index in SON under the RCP8.5 scenario. Shown are (a) 850-hPa H_e (gpm K^{-1}), (b) 850-hPa ζ ($10^{-6} \text{ s}^{-1} \text{ K}^{-1}$), (c) precipitation ($\text{mm day}^{-1} \text{ K}^{-1}$), (d) vertical velocity at 500 hPa ($10^{-2} \text{ Pa s}^{-1} \text{ K}^{-1}$), (e) vertical moisture advection ($\text{mm day}^{-1} \text{ K}^{-1}$), and (f) SST (K K^{-1}). The significant changes that agree in sign by more than 70% of the individual models are stippled. In (b)–(f), the contours indicate the anomalies during El Niño developing in autumn in 20C simulated by the Historical experiment, where the thick (dashed) contours indicate zero (negative) values.

21C, especially off the southwest coast of Sumatra (shading in Fig. 3f). Meanwhile, the positive SSTA over the western TIO is also slightly weakened, which further weakens the zonal gradient of SSTA across TIO basin. Previous studies suggested that a positive IOD usually occurs during the El Niño developing phase (Li et al. 2003; Yamagata et al. 2004), and this feature is captured by the MMM of the CMIP5 models (Fig. S3). The anomalous IOD index, defined as the zonal gradient of SSTA (Saji et al. 1999), is weakened by 14% and 21% K^{-1} under the RCP8.5 and RCP4.5 scenarios, respectively, which agrees with over 70% of the individual models (Fig. 4). Although it is proposed that the overall amplitude of IOD variability remains generally unchanged with large uncertainty under global warming (Zheng et al. 2013; Hui and Zheng 2018), our results

imply that the impact of El Niño on IOD may become weaker under a warmer climate.

The intermodel relationship between the change in precipitation anomaly averaged over MC (P index) and the change in atmospheric circulation anomaly over TIO (U index and ζ index) is examined in Fig. 5. First, it is clear that the models with a stronger reduction in the precipitation anomaly over MC are associated with a stronger reduction in the atmospheric circulation anomaly over TIO, including the equatorial easterly wind anomaly (Fig. 5a) and the anomalous anticyclones straddling the equator (Fig. 5b). Second, even if the precipitation anomaly over MC changes little in the MMM due to the compensation of increased moisture and weakened anomalous subsidence, the atmospheric circulation anomaly over TIO

FIG. 4. Percentage changes in the anomalous IOD index in the autumn of the developing El Niño under the RCP8.5 and RCP4.5 scenarios scaled by the amplitude of tropical-mean SST warming ($\% \text{ K}^{-1}$). The thin bar shows the intermodel uncertainty indicated by the 30th and 70th percentile of individual models.

is significantly weakened as evidenced by more than 70% of the individual models, regardless of the circulation index (Figs. 5a,b). The above two features are also evident in RCP4.5 scenario and depend little on the scenario (Figs. 5c,d). The intermodel correlation coefficients range from 0.69 to 0.76 for these two circulation indices under both scenarios (Figs. 5a–d), suggesting that the latent heating anomaly over MC explains about half of the intermodel variance of the change in the atmospheric circulation anomaly over TIO.

Based on the CGCM projections under global warming scenarios, the atmospheric circulation anomaly over TIO stimulated by the developing El Niño is substantially weakened under a warmer climate, even if the strength of El Niño remains unchanged. The model results are consistent with the observation in recent decades that the impact of the developing El Niño on TIO climate variability and South Asian monsoon rainfall is weakened (Kumar et al. 1999; Cash et al. 2017; Ham et al. 2017). The regression method adopted here implies an unchanged amplitude of variability of Niño-3.4 index, but the weakened zonal SSTA gradient across the TIO basin may contribute to the weakened atmospheric circulation anomaly over TIO. It is unclear whether the change in SSTA is the primary cause for the weakened atmospheric circulation anomaly, and idealized AGCM simulations under

increased mean-state SST with unchanged interannual variability of SSTA are analyzed in the next subsection to address this question.

b. Understanding the projected changes by AGCM simulations

To investigate how the anomalous atmospheric circulation will change in a warmer climate under the assumption that the El Niño-related SSTA remains unchanged, AMIP4K and AMIPFuture experiments performed by 11 models are compared with their AMIP experiments. Although the observed SST used in the AMIP experiment is a result of air–sea interaction, the interannual variability over tropical regions can be well reproduced by forcing AGCMs with observed SST (Gates et al. 1999; Zhou et al. 2018). In both AMIP4K and AMIPFuture experiments, the interannual variability component of global SST is exactly the same as in AMIP experiment, but a mean-state SST warming is employed. The MMM of the change in the atmospheric circulation anomalies over TIO during El Niño developing in autumn in AMIP4K and AMIPFuture experiments are shown in Fig. 6.

In both AMIP4K and AMIPFuture experiments, a substantially weakened atmospheric circulation anomaly over TIO is seen, evidenced by the weakened easterly wind anomaly along equatorial TIO and weakened anomalous zonal H_e gradient (Figs. 6a,b). The anomalous anticyclones on both side of the equator are also weakened, associated with weakened negative (positive) ζ anomaly over the northern (southern) TIO (Figs. 6c,d). The change in precipitation anomaly around MC is ambiguous and insignificant in both AMIP4K and AMIPFuture experiments (Figs. 6e,f), suggesting that the atmospheric circulation anomaly over TIO is still weakened even if the latent heating anomaly over MC is generally unchanged. Similar as in the CGCMs, the amplitude of change in the precipitation over MC explains a substantial fraction of the intermodel spread in the changes of atmospheric circulation over TIO, although the positive correlation between the P index and ζ index is slightly below the 95% confidence level in the AMIP4K experiment (Fig. S4). The similar results between CGCMs and AGCMs suggest the primary cause for the weakened atmospheric circulation anomaly over TIO is not the changed SSTA, despite the warming pattern of mean-state SST.

To quantitatively evaluate the changes, Fig. 7 shows the changes in the anomalies of the P index, U index, and ζ index for RCP8.5, RCP4.5, AMIP4K, and AMIPFuture experiments, in terms of the percentage changes relative to 20C scaled by tropical-mean SST warming. The MMM of the change in precipitation

FIG. 5. The projected changes in the regressed anomalies of regional averaged quantities onto the Niño-3.4 index in SON by the individual models. (a),(c) The change in the U index anomaly as a function of the change in the P index anomaly. (b),(d) The change in the ζ index anomaly as a function of the change in the P index anomaly. Here, (a) and (b) are based on the RCP8.5 scenario, while (c) and (d) are based on the RCP4.5 scenario. The blue dots indicate the individual models, and the red circle indicates the MMM. The intermodel correlation coefficients are marked in the upper-right corner of each panel, and the correlation coefficients significant at the 95% confidence level are marked with an asterisk.

anomaly over MC is weak and insignificant (blue bars in Fig. 7), possibly due to the compensation between increased mean-state moisture availability and weakened anomalous subsidence. Although the latent heating anomaly over MC is almost unchanged, the weakened atmospheric circulation anomaly over TIO is significant or marginally significant in all of the four experiments, in terms of both the U index (red bars in Fig. 7) and the ζ index (orange bars in Fig. 7). The amplitude of reduction in the TIO circulation anomaly is about 12%–13% K^{-1} in the CGCM in terms of both

the U index and ζ index, regardless of the scenario, but it is only about 8% K^{-1} in AMIP4K and AMIPFuture experiments. The amplitude of reduction is higher in CGCMs than in AGCMs, suggesting the change in SSTA in the CGCMs may further weaken the atmospheric circulation anomaly.

The comparison between AGCM and CGCM experiments suggests that the air–sea interaction is not the primary cause for the weakened TIO circulation anomaly during the developing El Niño, but the feedback between atmospheric circulation anomaly and the

FIG. 6. The MMM-simulated changes of the regressed atmospheric anomalies onto the Niño-3.4 index in SON in the (left) AMIP4K and (right) AMIPFuture experiments, compared to the AMIP experiment by the AGCMs. Shown are (a),(b) 850-hPa wind and H_e' ; (c),(d) 850-hPa ζ ; and (e),(f) precipitation. The contours indicate the anomalies during El Niño developing in autumn in 20C simulated by the AMIP experiment, where the thick (dashed) contours indicate zero (negative) values.

zonal gradient of SSTa over TIO acts to further weaken the circulation anomaly over TIO. Although the negative diabatic heating anomaly over MC directly drives the atmospheric circulation anomaly over TIO, the atmospheric circulation anomaly over TIO is substantially weakened despite an insignificant change in the diabatic heating anomaly over MC in both CGCM and AGCM experiments. It seems that a latent heating anomaly over MC could only stimulate a weaker atmospheric circulation anomaly over TIO during the developing El Niño under a warmer climate. Previous studies claimed that the mean-state static stability is robustly enhanced over

the entire tropics under global warming (Knutson and Manabe 1995; Schneider et al. 2010), and it acts to weaken the atmospheric circulation stimulated by a diabatic heat source (Schneider et al. 2010; Ma et al. 2012; Li et al. 2015; Qu and Huang 2016; He et al. 2017). Therefore, the possible role of enhanced tropical static stability is examined in the next section.

c. Understanding the role of enhanced static stability by LBM simulations

The heating anomaly over MC directly drives the atmospheric circulation anomalies over TIO during the

FIG. 7. The MMM of the percentage changes in the anomalies of the P index (blue bar), U index (red bar), and ζ index (orange bar) in the experiments performed by CGCMs and AGCMs, scaled by tropical-mean SST warming ($\% \text{ K}^{-1}$). The thin black bar shows the uncertainty range indicated by the 30% and 70% percentiles of the individual models.

El Niño developing phase via a Gill-type response, which is evident in both CGCMs and AGCMs (Fig. 1). As the change in latent heating anomaly over MC is weak and insignificant in both CGCM and AGCM simulations but the mean-state static stability are robustly enhanced under global warming (Schneider et al. 2010), LBM is used to examine the impact of enhanced static stability on the atmospheric circulation anomaly stimulated by a fixed heating anomaly over MC. In the control experiment (S20C), the mean state is derived from the 1950–99 period of the tNCEP–NCAR reanalysis in SON. In another experiment (S21C), the mean state of the NCEP–NCAR reanalysis is modified by adding a vertical profile of global averaged warming (Fig. S5a) to temperature at each grid point, to mimic the increased tropospheric static stability under RCP8.5. The LBM is forced by the regressed anomaly of observed three-dimensional diabatic heating over the MC region (10°S – 10°N , 80° – 120°E) onto the Niño-3.4 index (Fig. S5b shows its vertical profile), and the heating anomaly is exactly the same for the S20C and S21C experiments. Each experiment is run for 50 days to obtain a steady response, and the averaged anomalies for the last 20 days are shown.

Forced by the negative heating anomaly over MC, an easterly wind anomaly appears over the equatorial Indian Ocean, associated with a pair of anticyclone anomalies over the northern Indian Ocean and southern Indian Ocean on both sides of the equator (Fig. 8a), reminiscent of the atmospheric circulation anomalies during the El Niño developing in autumn in observation and CMIP5 models (Fig. 1). If the static stability of the troposphere is enhanced, similar atmospheric circulation anomaly is stimulated by the heating anomaly over MC but it is weaker (Fig. 8b). The difference between the S21C and S20C experiments is characterized by a westerly wind anomaly along equatorial Indian Ocean,

FIG. 8. The LBM simulated response of 850-hPa wind (vectors) and ζ (shading; 10^{-6} s^{-1}) to the negative heating over MC, and its sensitivity to mean-state static stability. (a) The atmospheric circulation anomalies stimulated by a negative heating anomaly over MC under the 20C mean state derived from the NCEP–NCAR reanalysis. (b) As in (a), but under a mean state with higher static stability, where the mean state is obtained by adding the global averaged vertical profile of warming under RCP8.5 to NCEP–NCAR data. (c) The difference of the atmospheric circulation anomalies between (b) and (a) scaled by 2.49 K (i.e., the amplitude of tropical SST warming under RCP8.5).

associated with positive (negative) ζ anomaly over TIO on the northern (southern) side of the equator (Fig. 8c), which is opposite to the atmospheric circulation anomaly in the S20C experiment. Quantitatively, the atmospheric circulation anomaly over TIO becomes weaker by about $6\% \text{ K}^{-1}$ in terms of both the U index and ζ index, quite close to the amplitude of reduction in the AMIP4K and AMIPFuture experiments. Therefore, the increased atmospheric static stability alone results in a

weakened impact of the developing El Niño on the atmospheric circulation anomaly over TIO, regardless of any air–sea interaction or any change in the atmospheric heating anomaly.

Based on the comparison among a hierarchy of model experiments, the primary cause for the weakened atmospheric circulation anomaly over TIO during the developing El Niño is probably the enhanced tropospheric static stability, and a secondary cause is the air–sea interaction within TIO. As the static stability of the atmosphere increases under a warming climate (Schneider et al. 2010; Ma et al. 2012; Li et al. 2015), the atmospheric circulation anomaly over TIO becomes less sensitive to the El Niño–induced negative heating anomaly over MC, and the negative heating anomaly over MC stimulates a weaker atmospheric circulation anomaly over TIO. The weakened atmospheric circulation anomaly over TIO further reduces the zonal gradient of SSTA over TIO via local air–sea interaction (Saji et al. 1999; Li et al. 2003; Yamagata et al. 2004), and further weaken the atmospheric circulation anomaly over TIO.

5. Conclusions and discussion

a. Conclusions

In this study, the possible change of the TIO climate variability associated with the developing El Niño is investigated. Both observation and models consistently show that the negative latent heating anomaly around MC is directly responsible for the atmospheric circulation anomaly over TIO during El Niño developing in autumn. The RCP8.5 and RCP4.5 experiments of 30 CGCMs are compared with their Historical experiment, to investigate the possible response of the El Niño–induced TIO climate variability to global warming. AMIP-type experiments performed by 11 CMIP5 models are also adopted, to investigate whether the El Niño–induced TIO climate variability is weakened if the El Niño–related SSTA remains unchanged under the mean-state warming. Finally, LBM is adopted to examine the possible impact of the enhanced tropospheric static stability on the anomalous atmospheric circulation stimulated by a diabatic heating anomaly over MC. The major findings are summarized as follows:

- 1) As consistently projected by CGCMs under the RCP8.5 and RCP4.5 scenarios, the impact of the developing El Niño on the anomalous atmospheric circulation over TIO is substantially weakened by about 12%–13% K^{-1} under global warming, and the associated zonal gradient of SSTA across TIO basin is also weakened. The latent heating anomaly

over MC explains about half of the intermodel uncertainty of the change in atmospheric circulation anomaly over TIO, but the MMM-projected change in latent heating anomaly over MC is insignificant and it cannot explain the overall weakening of TIO circulation anomaly in the MMM. Based on both AMIP4K and AMIPFuture experiments compared with the AMIP experiment, the atmospheric circulation anomaly over TIO during El Niño developing in autumn is robustly weakened by about 8% K^{-1} even if the SSTA associated with El Niño remains exactly unchanged, independent of the warming pattern of mean-state SST. Therefore, the weakened zonal gradient of SSTA across TIO basin in CGCMs is not the primary cause for the weakened TIO circulation anomaly.

- 2) The enhanced mean-state tropospheric static stability over the tropics is a robust phenomenon under global warming, and LBM experiments suggest that the response of the anomalous atmospheric circulation over TIO to the diabatic heating anomaly over MC is substantially weakened by about 6% K^{-1} as the mean-state static stability increases, despite no change in the heating anomaly over MC. The enhanced tropospheric static stability is probably the primary cause for the weakened atmospheric circulation anomaly over TIO during the El Niño developing phase, and the air–sea interaction over TIO and the associated reduction in zonal SSTA gradient across TIO basin is probably a secondary cause. Under a warmer climate, the atmospheric circulation anomaly over TIO stimulated by the heating anomaly over MC is weakened due to the enhanced tropospheric static stability, and the weakened atmospheric circulation anomaly over TIO results in a weakened zonal gradient of SSTA across the TIO basin, and further acts to weaken the anomalous atmospheric circulation over TIO through local air–sea interaction.

b. Discussion

Although previous studies proposed an enhanced impact of extreme ENSO on the atmospheric circulation anomalies and the associated teleconnection (Cai et al. 2014, 2015), our results suggest that the enhanced atmospheric static stability is a nonnegligible factor that acts to weaken the atmospheric circulation anomalies associated with ENSO. The enhanced static stability is a robust and spatially uniform phenomenon around the entire tropics (Knutson and Manabe 1995; Schneider et al. 2010). It explains the weakened atmospheric circulation anomalies over TIO stimulated by a diabatic

heating anomaly over MC, but it cannot explain the intermodel uncertainty in the changes of atmospheric circulation anomalies (Table S2). The enhanced static stability may also act to weaken the ENSO-related atmospheric circulation anomalies over other tropical oceans, but the mechanism for the response of ENSO-related atmospheric circulation anomalies in midlatitudes may be more complex due to the synoptic transients (Coumou et al. 2015; Lu et al. 2017), which deserves further study.

Air–sea interaction over TIO plays a key role in the interannual variability associated with ENSO (Li et al. 2003; Yamagata et al. 2004; Feng and Duan 2018). Under a warming climate, both the atmospheric circulation and SST anomalies over TIO are changed, and it is hard to distinguish the cause and effect. Based on the AMIP, AMIP4K, and AMIPFuture experiments of AGCMs and its comparison with the CGCM projections, this study suggested that the weakened easterly wind anomalies over TIO due to enhanced static stability may be responsible for the weakened positive IOD condition. But a clear limitation is that the lack of air–sea interaction in the AGCM may bias the result. Further study is needed to address how the bias in the simulation of 20C climate leads to the bias in the response to global warming (Xie et al. 2015; Eyring et al. 2019) before removing the systematic model bias.

Acknowledgments. The authors wish to acknowledge Prof. Masahiro Watanabe for offering the code of LBM. This work was supported by the National Key Research and Development Program of China (2017YFA0604601), and the National Natural Science Foundation of China (41875081), and Open Research Fund Program of Key Laboratory of Meteorological Disaster of Ministry of Education (KLME1601).

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).
- Cai, W., and Coauthors, 2014: Increasing frequency of extreme El Niño events due to greenhouse warming. *Nat. Climate Change*, **4**, 111–116, <https://doi.org/10.1038/nclimate2100>.
- , and Coauthors, 2015: Increased frequency of extreme La Niña events under greenhouse warming. *Nat. Climate Change*, **5**, 132–137, <https://doi.org/10.1038/nclimate2492>.
- Cash, B. A., and Coauthors, 2017: Sampling variability and the changing ENSO–monsoon relationship. *Climate Dyn.*, **48**, 4071–4079, <https://doi.org/10.1007/s00382-016-3320-3>.
- Chen, L., T. Li, and Y. Yu, 2015: Causes of strengthening and weakening of ENSO amplitude under global warming in four CMIP5 models. *J. Climate*, **28**, 3250–3274, <https://doi.org/10.1175/JCLI-D-14-00439.1>.
- , —, —, and S. K. Behera, 2017: A possible explanation for the divergent projection of ENSO amplitude change under global warming. *Climate Dyn.*, **49**, 3799–3811, <https://doi.org/10.1007/s00382-017-3544-x>.
- Cherchi, A., and A. Navarra, 2013: Influence of ENSO and of the Indian Ocean dipole on the Indian summer monsoon variability. *Climate Dyn.*, **41**, 81–103, <https://doi.org/10.1007/s00382-012-1602-y>.
- Collins, M., and Coauthors, 2010: The impact of global warming on the tropical Pacific Ocean and El Niño. *Nat. Geosci.*, **3**, 391–397, <https://doi.org/10.1038/ngeo868>.
- Coumou, D., J. Lehmann, and J. Beckmann, 2015: The weakening summer circulation in the Northern Hemisphere midlatitudes. *Science*, **348**, 324–327, <https://doi.org/10.1126/science.1261768>.
- Deser, C., A. Phillips, V. Bourdette, and H. Teng, 2012: Uncertainty in climate change projections: the role of internal variability. *Climate Dyn.*, **38**, 527–546, <https://doi.org/10.1007/s00382-010-0977-x>.
- Eyring, V., and Coauthors, 2019: Taking climate model evaluation to the next level. *Nat. Climate Change*, **9**, 102–110, <https://doi.org/10.1038/s41558-018-0355-y>.
- Fan, L., Q. Liu, C. Wang, and F. Guo, 2017: Indian Ocean dipole modes associated with different types of ENSO development. *J. Climate*, **30**, 2233–2249, <https://doi.org/10.1175/JCLI-D-16-0426.1>.
- Feng, R., and W. Duan, 2018: The role of initial signals in the tropical Pacific Ocean in predictions of negative Indian Ocean dipole events. *Sci. China Earth Sci.*, **61**, 1832–1843, <https://doi.org/10.1007/s11430-018-9296-2>.
- Gates, W. L., and Coauthors, 1999: An overview of the results of the Atmospheric Model Intercomparison Project (AMIP I). *Bull. Amer. Meteor. Soc.*, **80**, 29–55, [https://doi.org/10.1175/1520-0477\(1999\)080<0029:AOTRO>2.0.CO;2](https://doi.org/10.1175/1520-0477(1999)080<0029:AOTRO>2.0.CO;2).
- 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.4971064905>.
- Gleckler, P. J., K. E. Taylor, and C. Doutriaux, 2008: Performance metrics for climate models. *J. Geophys. Res.*, **113**, D06104, <https://doi.org/10.1029/2007JD008972>.
- Ham, Y.-G., J.-Y. Choi, and J.-S. Kug, 2017: The weakening of the ENSO–Indian Ocean dipole (IOD) coupling strength in recent decades. *Climate Dyn.*, **49**, 249–261, <https://doi.org/10.1007/s00382-016-3339-5>.
- He, C., and T. Li, 2019: Does global warming amplify interannual climate variability? *Climate Dyn.*, **52**, 2667–2684, <https://doi.org/10.1007/s00382-018-4286-0>.
- , B. Wu, L. Zou, and T. Zhou, 2017: Responses of the summertime subtropical anticyclones to global warming. *J. Climate*, **30**, 6465–6479, <https://doi.org/10.1175/JCLI-D-16-0529.1>.
- , A. Lin, D. Gu, C. Li, B. Zheng, B. Wu, and T. Zhou, 2018: Using eddy geopotential height to measure the western North Pacific subtropical high in a warming climate. *Theor. Appl. Climatol.*, **131**, 681–691, <https://doi.org/10.1007/s00704-016-2001-9>.
- , T. Zhou, and T. Li, 2019: Weakened anomalous western North Pacific anticyclone during an El Niño–decaying summer under a warmer climate: Dominant role of the weakened impact of the tropical Indian Ocean on the atmosphere. *J. Climate*, **32**, 213–230, <https://doi.org/10.1175/JCLI-D-18-0033.1>.
- Hu, H., Q. Wu, and Z. Wu, 2018: Influences of two types of El Niño event on the northwest Pacific and tropical Indian Ocean SST anomalies. *J. Oceanol. Limnol.*, **36**, 33–47, <https://doi.org/10.1007/s00343-018-6296-5>.

- Huang, B., and Coauthors, 2017: Extended Reconstructed Sea Surface Temperature, version 5 (ERSSTv5): Upgrades, validations, and intercomparisons. *J. Climate*, **30**, 8179–8205, <https://doi.org/10.1175/JCLI-D-16-0836.1>.
- Hui, C., and X.-T. Zheng, 2018: Uncertainty in Indian Ocean dipole response to global warming: The role of internal variability. *Climate Dyn.*, **51**, 3597–3611, <https://doi.org/10.1007/s00382-018-4098-2>.
- Izumo, T., M. Lengaigne, J. Vialard, J.-J. Luo, T. Yamagata, and G. Madec, 2014: Influence of Indian Ocean dipole and Pacific recharge on following year's El Niño: Interdecadal robustness. *Climate Dyn.*, **42**, 291–310, <https://doi.org/10.1007/S00382-012-1628-1>.
- Jiang, W., G. Huang, P. Huang, and K. Hu, 2018: Weakening of northwest Pacific anticyclone anomalies during post-El Niño summers under global warming. *J. Climate*, **31**, 3539–3555, <https://doi.org/10.1175/JCLI-D-17-0613.1>.
- Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471, [https://doi.org/10.1175/1520-0477\(1996\)077<0437:TNYRP>2.0.CO;2](https://doi.org/10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2).
- Knutson, T. R., and S. Manabe, 1995: Time-mean response over the tropical Pacific to Increased CO₂ in a coupled ocean–atmosphere model. *J. Climate*, **8**, 2181–2199, [https://doi.org/10.1175/1520-0442\(1995\)008<2181:TMROTT>2.0.CO;2](https://doi.org/10.1175/1520-0442(1995)008<2181:TMROTT>2.0.CO;2).
- Kripalani, R. H., J. H. Oh, and H. S. Chaudhari, 2010: Delayed influence of the Indian Ocean dipole mode on the East Asia–west Pacific monsoon: Possible mechanism. *Int. J. Climatol.*, **30**, 197–209, <https://doi.org/10.1002/JOC.1890>.
- Kug, J.-S., T. Li, S.-I. An, I.-S. Kang, J.-J. Luo, S. Masson, and T. Yamagata, 2006: Role of the ENSO–Indian Ocean coupling on ENSO variability in a coupled GCM. *Geophys. Res. Lett.*, **33**, L09710, <https://doi.org/10.1029/2005GL024916>.
- Kumar, K. K., B. Rajagopalan, and M. A. Cane, 1999: On the weakening relationship between the Indian monsoon and ENSO. *Science*, **284**, 2156–2159, <https://doi.org/10.1126/science.284.5423.2156>.
- Lau, N. C., and M. J. Nath, 2003: Atmosphere–ocean variations in the Indo-Pacific sector during ENSO episodes. *J. Climate*, **16**, 3–20, [https://doi.org/10.1175/1520-0442\(2003\)016<0003:AOVITI>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<0003:AOVITI>2.0.CO;2).
- Li, T., B. Wang, C. P. Chang, and Y. S. Zhang, 2003: A theory for the Indian Ocean dipole–zonal mode. *J. Atmos. Sci.*, **60**, 2119–2135, [https://doi.org/10.1175/1520-0469\(2003\)060<2119:ATFTIO>2.0.CO;2](https://doi.org/10.1175/1520-0469(2003)060<2119:ATFTIO>2.0.CO;2).
- , L. Zhang, and H. Murakami, 2015: Strengthening of the Walker circulation under global warming in an aqua-planet general circulation model simulation. *Adv. Atmos. Sci.*, **32**, 1473–1480, <https://doi.org/10.1007/s00376-015-5033-7>.
- Liu, L., G. Yang, X. Zhao, L. Feng, G. Han, Y. Wu, and W. Yu, 2017: Why was the Indian Ocean dipole weak in the context of the extreme El Niño in 2015? *J. Climate*, **30**, 4755–4761, <https://doi.org/10.1175/JCLI-D-16-0281.1>.
- Lu, J., K. Sakaguchi, Q. Yang, L. R. Leung, G. Chen, C. Zhao, E. Swenson, and Z. J. Hou, 2017: Examining the hydrological variations in an aquaplanet world using wave activity transformation. *J. Climate*, **30**, 2559–2576, <https://doi.org/10.1175/JCLI-D-16-0561.1>.
- Luo, J. J., R. C. Zhang, S. K. Behera, Y. Masumoto, F. F. Jin, R. Lukas, and T. Yamagata, 2010: Interaction between El Niño and extreme Indian Ocean dipole. *J. Climate*, **23**, 726–742, <https://doi.org/10.1175/2009JCLI1304.1>.
- Ma, J., S.-P. Xie, and Y. Kosaka, 2012: Mechanisms for tropical tropospheric circulation change in response to global warming. *J. Climate*, **25**, 2979–2994, <https://doi.org/10.1175/JCLI-D-11-00048.1>.
- Maher, N., D. Matei, S. Milinski, and J. Marotzke, 2018: ENSO change in climate projections: Forced response or internal variability? *Geophys. Res. Lett.*, **45**, 11 390–11 398, <https://doi.org/10.1029/2018GL079764>.
- Ni, Y., and P.-C. Hsu, 2018: Inter-annual variability of global monsoon precipitation in present-day and future warming scenarios based on 33 Coupled Model Intercomparison Project Phase 5 models. *Int. J. Climatol.*, **38**, 4875–4890, <https://doi.org/10.1002/joc.5704>.
- Power, S. B., F. Delage, R. Colman, and A. Moise, 2012: Consensus on twenty-first-century rainfall projections in climate models more widespread than previously thought. *J. Climate*, **25**, 3792–3809, <https://doi.org/10.1175/JCLI-D-11-00354.1>.
- , —, C. Chung, G. Kociuba, and K. Keay, 2013: Robust twenty-first-century projections of El Niño and related precipitation variability. *Nature*, **502**, 541–545, <https://doi.org/10.1038/nature12580>.
- Qu, X., and G. Huang, 2016: The global warming-induced South Asian high change and its uncertainty. *J. Climate*, **29**, 2259–2273, <https://doi.org/10.1175/JCLI-D-15-0638.1>.
- Saji, N. H., B. N. Goswami, P. N. Vinayachandran, and T. Yamagata, 1999: A dipole mode in the tropical Indian Ocean. *Nature*, **401**, 360–363, <https://doi.org/10.1038/43854>.
- Schneider, T., P. A. O’Gorman, and X. J. Levine, 2010: Water vapor and the dynamics of climate changes. *Rev. Geophys.*, **48**, RG3001, <https://doi.org/10.1029/2009RG000302>.
- Taylor, K. E., R. J. Stouffer, and G. A. Meehl, 2012: An overview of CMIP5 and the experiment design. *Bull. Amer. Meteor. Soc.*, **93**, 485–498, <https://doi.org/10.1175/BAMS-D-11-00094.1>.
- Ummenhofer, C. C., A. Sen Gupta, Y. Li, A. S. Taschetto, and M. H. England, 2011: Multi-decadal modulation of the El Niño–Indian monsoon relationship by Indian Ocean variability. *Environ. Res. Lett.*, **6**, 034006, <https://doi.org/10.1088/1748-9326/6/3/034006>.
- Van Vuuren, D., and Coauthors, 2011: The representative concentration pathways: An overview. *Climatic Change*, **109**, 5–31, <https://doi.org/10.1007/s10584-011-0148-z>.
- Wang, L., A. Deng, and R. Huang, 2019: Wintertime internal climate variability over Eurasia in the CESM large ensemble. *Climate Dyn.*, **52**, 6735–6748, <https://doi.org/10.1007/S00382-018-4542-3>.
- Wang, X., and C. Wang, 2014: Different impacts of various El Niño events on the Indian Ocean dipole. *Climate Dyn.*, **42**, 991–1005, <https://doi.org/10.1007/s00382-013-1711-2>.
- Watanabe, M., and M. Kimoto, 2000: Atmosphere–ocean thermal coupling in the North Atlantic: A positive feedback. *Quart. J. Roy. Meteor. Soc.*, **126**, 3343–3369, <https://doi.org/10.1002/qj.49712657017>.
- Xavier, P. K., C. Marzin, and B. N. Goswami, 2007: An objective definition of the Indian summer monsoon season and a new perspective on the ENSO–monsoon relationship. *Quart. J. Roy. Meteor. Soc.*, **133**, 749–764, <https://doi.org/10.1002/qj.45>.
- Xie, S.-P., and Coauthors, 2015: Towards predictive understanding of regional climate change. *Nat. Climate Change*, **5**, 921–930, <https://doi.org/10.1038/nclimate2689>.
- Xu, K., C. Zhu, and W. Wang, 2016: The cooperative impacts of the El Niño–Southern Oscillation and the Indian Ocean dipole on the interannual variability of autumn rainfall in China. *Int. J. Climatol.*, **36**, 1987–1999, <https://doi.org/10.1002/joc.4475>.

- Yamagata, T., S. K. Behera, J. J. Luo, S. Masson, M. R. Jury, and S. A. Rao, 2004: Coupled ocean–atmosphere variability in the tropical Indian Ocean. *Earth’s Climate: The Ocean–Atmosphere Interaction, Geophys. Monogr.*, Vol. 147, Amer. Geophys. Union, 189–211, <https://doi.org/10.1029/147GM12>.
- Yanai, M., and T. Tomita, 1998: Seasonal and interannual variability of atmospheric heat sources and moisture sinks as determined from NCEP–NCAR reanalysis. *J. Climate*, **11**, 463–482, [https://doi.org/10.1175/1520-0442\(1998\)011<0463:SAIVOA>2.0.CO;2](https://doi.org/10.1175/1520-0442(1998)011<0463:SAIVOA>2.0.CO;2).
- Yang, S., Z. Li, J.-Y. Yu, X. Hu, W. Dong, and S. He, 2018: El Niño–Southern Oscillation and its impact in the changing climate. *Natl. Sci. Rev.*, **5**, 840–857, <https://doi.org/10.1093/nsr/nwy046>.
- Yuan, Y., and C. Li, 2008: Decadal variability of the IOD-ENSO relationship. *Chin. Sci. Bull.*, **53**, 1745–1752, <https://doi.org/10.1007/S11434-008-0196-6>.
- Zheng, X.-T., S.-P. Xie, Y. Du, L. Liu, G. Huang, and Q. Liu, 2013: Indian Ocean dipole response to global warming in the CMIP5 multimodel ensemble. *J. Climate*, **26**, 6067–6080, <https://doi.org/10.1175/JCLI-D-12-00638.1>.
- Zhou, Z.-Q., S.-P. Xie, X.-T. Zheng, Q. Liu, and H. Wang, 2014: Global warming–induced changes in El Niño teleconnections over the North Pacific and North America. *J. Climate*, **27**, 9050–9064, <https://doi.org/10.1175/JCLI-D-14-00254.1>.
- , —, G. J. Zhang, and W. Zhou, 2018: Evaluating AMIP skill in simulating interannual variability over the Indo–western Pacific. *J. Climate*, **31**, 2253–2265, <https://doi.org/10.1175/JCLI-D-17-0123.1>.